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1 FROM GONDWANA RIFTING TO ALPINE OROGENY:
2 DETRITAL ZIRCON GEOCHRONOLOGIC AND
3 PROVENANCE SIGNALS FROM THE KOPET DAGH
4 BASIN (NE IRAN)

5 ALI MOHAMMADI^{*,†}, JONAS B. RUH^{**}, MARCEL GUILLONG^{**}, OSCAR LAURENT^{***,****}, and
6
7 LOTFOLLAH AGHAJARI[§]

8 * Eurasia Institute of Earth Sciences, Istanbul Technical University, Maslak, Istanbul, Turkey

9 ** Department of Earth Sciences, ETH Zürich, Sonneggstrasse 5, 8092 Zürich, Switzerland

10 *** CNRS, Géosciences Environnement Toulouse, Observatoire Midi-Pyrénées, 14 avenue Edouard Belin F-
11 31400 Toulouse, France

12 § Exploration Directorate, National Iranian Oil Company, Sheikh Baha'i Square, Seoul Street, Tehran, Iran
13

14 **ABSTRACT.** The Kopet Dagh mountains in NE Iran exhibit a 7-km-thick continuous
15 sedimentary sequence recording detritus from exposed surrounding terranes from the last 175
16 Myr. This work presents a multi-disciplinary geochronologic and provenance analysis in an
17 attempt to identify and date major geologic events along the northern segment of the Tethys
18 and reconstruct the regional tectonic history from Gondwana-related rifting until Alpine
19 orogeny. Sandstone framework, heavy mineral analysis, U-Pb dating of detrital zircons, and
20 Hf-isotope ratio measurements on dated zircons from Triassic to Paleocene sandstones indicate
21 three main tectonic events that include early Silurian intracontinental rifting (opening of Paleo-
22 Tethys), early Carboniferous rifting of a back-arc basin (Aghdarband Basin), and Late Triassic
23 collisional to post-collisional magmatism (Paleo-Tethys collision). Mineralogical and age peak
24 considerations indicate that detritus was supplied from the south into the extensional Kopet
25 Dagh Basin during Middle Jurassic, while Cretaceous to Paleocene sandstones show signs of
26 increasing cannibalism.

27 **Keywords:** Gondwana, Kopet Dagh Basin, Paleo-Tethys, Neo-Tethys, Rifting, Collision,
28 Detrital zircon, Hf isotopes, Provenance analysis
29

30 **INTRODUCTION**

31 The long-standing tectonic history between Gondwana and Eurasia includes several phases of
32 rifting and collision, recorded by magmatic and metamorphic episodes in various domains of
33 related orogenic belts; from the Alps in the west to the Himalayas in the east (for example

34 [Şengör, 1990a; 1990b; Stampfli, 2000; Stampfli and Borel, 2002](#)). The timing of main tectonic
35 episodes including rifting, subduction initiation, and collision are crucial for the reconstruction
36 of plate tectonics ([for example Stampfli and Borel, 2002](#)). Geochronology and geochemistry of
37 magmatic and metamorphic rocks are widely applied in geodynamic studies to constrain the
38 timing and other characteristics of tectonic events ([for example Barbarin, 1999; Brown, 2014](#)).
39 However, the occurrence and surface exposure of such igneous and metamorphic terranes may
40 be limited in some tectono-sedimentary domains because of complete erosion or coverage by
41 younger sediments. In addition, the application of different dating methods on igneous and
42 metamorphic rocks provides a wide range of ages, which makes the exact timing of tectonic
43 events difficult. In fact, timing constraints for tectonic events based only on magmatic or
44 metamorphic rocks may be misleading. A more accurate detection of large-scale tectonic events
45 can be found in clastic sedimentary successions, where the detritus source rapidly responds to
46 any plate tectonic reorganization ([Cawood and others, 2012; Fedo and others, 2003](#)). Therefore,
47 provenance analysis, zircon geochronology, and zircon isotopic composition on detrital
48 sandstones may help to improve the precision when constraining the timing of tectonic events
49 ([for example Cawood and others, 2012; Weltje and von Eynatten, 2004](#)).

50 Several episodes of continental rifting and collision took place between Gondwana and Eurasia
51 since the Palaeozoic until the Cenozoic that ultimately climaxed in the Alpine-Himalayan
52 orogeny ([for example Şengör and others, 1984; Stampfli, 2000; Stampfli and Borel, 2002;](#)
53 [Stöcklin, 1974](#)). Separation of several continental slivers and microcontinental blocks from
54 Gondwana occurred in two phases, coeval with the opening of two ocean basins, the Paleo-
55 Tethys and Neo-Tethys oceans ([for example Alavi, 1996; Stampfli, 2000; Şengör and others,](#)
56 [1984; Stöcklin, 1974](#)). The opening of the Iranian segment of the Paleo-Tethys between the
57 Iranian continent (Gondwana) and the Turan platform (Eurasia) was suggested to have occurred
58 in the Early Palaeozoic (Ordovician-Silurian) based on ages of oceanic basalts ([Boulin, 1988;](#)
59 [Stampfli, 2000](#)). The subsequent northwards subduction of the Paleo-Tethys beneath Eurasia

60 initiated during the Late Devonian and continued until the Late Triassic as shown from dating
61 of the Darrehanjir–Mashhad ophiolites (for example Moghadam and others, 2015a; Stampfli,
62 2000; Wilmsen and others, 2009). The opening of a new ocean (Neo-Tethys Ocean) between
63 the Arabian plate and the Central Iran block at the end of the Permian went hand in hand with
64 the northward movement of the Central Iran block, resulting in the closure of the Paleo-Tethys
65 Ocean in the Late Triassic, and the consequent collision of the Iranian plate with the Turan
66 block; the Cimmerian orogeny (for example Berberian and King, 1981; Muttoni and others,
67 2009; Şengör, 1979; Zanchetta and others, 2013). This Cimmerian orogeny prevented further
68 shortening along the northern segment of the Tethys and triggered the onset of subduction of
69 the southern Neo-Tethys Ocean (for example Zanchetta and others, 2013).

70 The northward subduction of the Neo-Tethys Ocean beneath the Central Iran block was
71 accompanied by the opening of the Greater Caucasus–South Caspian–Kopet Dagh basins
72 during the Middle Mesozoic, potentially in form of a larger back-arc basin (Alavi et al, 1997;
73 Brunet and Cloetingh, 2003; Zanchi and others, 2006). The Neo-Tethys Ocean finally closed
74 during the collision of the Arabian plate and Central Iran block in late Oligocene, which resulted
75 in the Alpine orogeny (Pirouz and others, 2017). This collision related to the closure of the Neo-
76 Tethys resulted in tectonic inversion of previously extensional domains (Greater Caucasus,
77 South Caspian, and Kopet Dagh basins) along the Paleo-Tethys suture (for example Robert and
78 others, 2014; Ruh and Vergés, 2018; Ruh, 2019).

79 The Kopet Dagh Basin contains a stratigraphic record covering the last ~175 Myr (Middle
80 Jurassic to Neogene sedimentary succession; Afshar Harb, 1982a; Moussavi-Harami and
81 Brenner, 1990). Detailed provenance analysis of this shallowing-upward (deep marine–marine–
82 deltaic–continental) sequence allows i) to reconstruct the Mesozoic-Tertiary sedimentary
83 evolution of the Kopet Dagh Basin, ii) to clarify the general tectonic setting associated with the
84 Turan and Iran plates, and finally iii) to precisely constrain the timing of rifting and collision
85 events related to the opening and closure of the Paleo-Tethys and Neo-Tethys oceans.

86 In this study, we present combined results from sandstone modal framework, heavy mineral
87 assemblage analysis, detrital zircon U-Pb age, and *in-situ* Hf isotope composition of Triassic
88 (pre-rift basement), and Middle Jurassic to Paleocene detrital sediments from the Kopet Dagh.
89 Results are interpreted to identify and characterize the respective source rocks in terms of ages
90 and genesis. Based on these data, we determine the provenance of sedimentary detritus and
91 reconstruct the variety of rift-related, continental arc, and post-collisional tectonic settings from
92 Gondwana-related rifting (opening of Paleo-Tethys) to Alpine orogeny (closure of Neo-
93 Tethys).

94

95 GOLOGY OF THE KOPET DAGH

96

Tectonic Setting

97 The Kopet Dagh mountains in NE Iran strike over 700 km along the Iran–Turkmenistan border
98 and form the northern limit of the Alpine-Himalayan orogeny (Golonka, 2004; Robert and
99 others, 2014; fig. 1). The Main Kopet Dagh Fault (MKF) and the East Alborz and Binalud
100 mountains mark the northern and the southern boundary of the Kopet Dagh Basin, respectively
101 (fig. 1). Tectonically, the Kopet Dagh Basin formed as an extensional back-arc to the Neo-
102 Tethys subduction system farther to the south (Mangino and Priestley, 1998). The onset of
103 extension is marked by a major nonconformity at the base, and rapid sediment facies and
104 thickness variations in the Middle Jurassic (165 Ma to 170 Ma) deep-marine turbiditic
105 sediments (Kavoosi and others, 2009; Robert and others, 2014; Thomas and others, 1999). In
106 contrast to the Caspian Sea Basin in the west, there is no evidence for the formation of oceanic
107 crust beneath the Kopet Dagh Basin (Brunet and others, 2003). After this rifting event, the basin
108 experienced a phase of tectonic quiescence for at least 130 Myr until Oligocene-Miocene, when
109 the extensional basins along the Paleo-Tethys suture experienced tectonic inversion as a
110 consequence of the collision of Arabian and Eurasian plates to form the Zagros mountain range,

111 and due to newly arranged relative plate motions (Hollingsworth and others, 2010; Lyberis and
112 Manby, 1999). This tectonic inversion is synchronous with the onset of deposition of
113 continental sedimentary facies in the Kopet Dagh Basin (Robertand others, 2014) and tectonic
114 uplift of the Alborz mountains (Razaeian and others, 2012). Uplift of the Kopet Dagh mountains
115 resulted in widespread folding and faulting of the Jurassic to Tertiary stratigraphy and an
116 estimated N-S shortening of ~75 km (Allen and others, 2003; Lyberis and Manby, 1999). The
117 compressional phase continued for about 15 Myr, with an increased component of dextral
118 strike-slip movement since ~10 Ma (Hollingsworth and others, 2006).

119

120

Stratigraphy

121 The study area is located in the eastern part of the Kopet Dagh Basin covered by the Darreh
122 Gaz, Mashhad, Sarakhs, Torbat-e-Heydarieh, and Torbat-e-Jam 1:250,000 geological maps
123 published by the Geological Survey of Iran (figs. 1, 2; Afshar 1982a, 1982b; Afshar and others,
124 1986; Behrouzi and others, 1993; Vaezi Pour and others, 1992). An up to seven-kilometers-
125 thick terrigenous and carbonate sedimentary succession was deposited into the Kopet Dagh
126 Basin between Middle Jurassic to Neogene (figs. 2, 3; Moussavi-Harami and Brenner, 1990).
127 The Middle Jurassic (late Bajocian to Bathonian) intervals of deep marine turbiditic
128 conglomerate, thin- and thick-bedded sandstones, siltstone, and mudstone (Kashafrud
129 Formation) are the oldest sediments deposited into the Kopet Dagh Basin (for example
130 Poursoltani and Gibling, 2011; Poursoltani, and others, 2007). The Middle Jurassic sediments
131 rest unconformably on Triassic basic intrusions and volcanogenic sedimentary rocks and
132 exhibit variations in thickness across major extensional basement faults, indicating a syn-rifting
133 depositional environment (Robert and others, 2014; Thomas and others, 1999). Thickness of
134 the turbiditic sediments vary between 280 m to over 2000 m in the East Kopet Dagh Basin (fig.
135 3). The Middle Jurassic turbiditic succession is unconformably overlain by Upper Jurassic
136 (Callovian-Kimmeridgian) thick (>1200 m) marine carbonates and shallow marine, coastal

137 plain siliciclastic sediments (Mozduran Formation; [Kavoosi and others, 2009](#); [Lasemi, 1995](#);
138 [fig. 3](#)). Thick fluvial siliciclastic sediments of the Early Cretaceous (Suriyeh Formation) record
139 epicontinental sea retrogradation in the eastern part of the Kopet Dagh Basin ([fig. 3](#); [Moussavi-
140 Harami and Brenner, 1990](#)). Contemporaneous deposition of marine carbonate indicates the
141 presence of a marine environment in the western Kopet Dagh Basin. Further transgression
142 established a marine depositional environment throughout the entire Kopet Dagh Basin until
143 Late Cretaceous (Turonian). This transgressions system is characterized by intervals of oolitic
144 limestone (Tirgan Formation), marl and shale (Sarcheshmeh Formation), shale (Sanganeh
145 Formation), and glauconitic-carbonatic sandstone and shale (Aitamir Formation; [fig. 3](#)). This
146 wide range of lithologies indicates sea level fluctuation and depositional environments
147 represented by tidal flats, lagoons, carbonate ramps, continental shelves, and shallow marine-
148 shorefaces. A short-lived sea retrogradation during the Late Cretaceous (Turonian) is responsible
149 for an unconformity between the shallow marine glauconitic shale (Aitamir Formation) and
150 shallow marine marl and limestones (Abderaz Formation; [fig. 3](#)). Between the Turonian until
151 the Paleocene, the East Kopet Dagh Basin experienced other transgressional episodes
152 characterized by marine environments including the deposition of shallow marine marl and
153 limestones (Abderaz Formation), shelf and open marine marl and shale (Abtalkh Formation),
154 shallow marine and shoreface glauconitic sandstone (Neyzar Formation), and shallow
155 carbonate platform sandy limestones (Kalat Formation; [fig. 3](#)). Basin-scale sea retrogradation
156 and surface exposure of strata is recorded by erosional surfaces within the uppermost part of
157 Late Cretaceous sandy limestones (Kalat Formation; [Mahboubi and others, 2006](#)). In early
158 Paleocene, reddish fluvial and lacustrine mudstones and sandstones (Pestehleigh Formation)
159 were unconformably deposited onto uppermost Cretaceous marine sediments. During the late
160 Paleocene, the Kopet Dagh Basin gradually changed from a shallow continental sedimentary
161 environment to a shallow carbonate platform with marl and limestone sediments (Chehel
162 Kaman Formation). The basin subsidence continued until Eocene-early Oligocene, when

163 marine shale and sandstones (Khangiran Formation) conformably overlaid late Paleocene
164 sediments. Repeated minor phases of transgressions and regressions followed into the Miocene,
165 until the basin was tectonically inverted during Alpine orogeny, accompanied by deposition of
166 continental conglomerates, sandstones, and shale (Neogene red beds; fig. 3).

167

168

METHODS

169 Nineteen sandstone samples including one sandstone sample of a lens within Triassic
170 polymictic conglomerate, and eighteen well-cemented medium-grained Jurassic to Paleocene
171 turbiditic, fluvial-deltaic, and continental sandstone samples were investigated (with prefix KD-
172 01–20 in the ETH collection). Studied sandstones represent five different litho-stratigraphic
173 formations (Qara-Gheitan, Kashafrud, Shurijeh, Aitamir, and Pestehleigh formations) of the
174 Kopet Dagh Basin (figs 1, 2, 3 and table S1). We combined sandstone framework composition
175 and heavy mineral analysis, detrital zircon U-Pb geochronology, and *in-situ* Hf-isotopic
176 composition to reconstruct the lithological composition of the source-rocks and the tectonic
177 evolution of the source area.

178

Sandstone thin section

179 Sandstones modal framework analysis was performed in thin sections stained for carbonates
180 and feldspars with the Gazzi-Dickinson method (Dickson, 1966; Norman, 1974). For statistical
181 reliability, at least 300 lithic fragments were counted in each thin section (Folk, 1980). The
182 Zuffa method was applied to count detrital grains (Zuffa, 1985). Sand size lithic fragments
183 (>63 µm) within rock fragments were counted as monomineralic grains. Results were
184 converted to percentages for compositional comparison (Weltje and von Eynatten, 2004; table
185 S2). Five standard complementary triangular diagrams were used to display the data (QFL,
186 QmFLt, LvLsLm, QmPK, and QpLvLsm; Folk, 1980; Dickinson, 1985).

187

188

Heavy mineral separation

189 Approximately 2-3 kg of fresh rock were collected at each sample site for heavy mineral
190 analysis. To obtain transparent heavy mineral fractions, sandstones were crushed with the
191 SelFrag apparatus batch equipment using high voltage (130-150 kV) pulse power technology,
192 which liberates morphologically intact minerals. From the <2 mm sieved fraction, carbonate
193 was dissolved in warm (60-70°C), 10%-acetic acid. Heavy minerals were pulled out in
194 separation funnels (Mange and Maurer, 1992) from 63 µm to 400 µm sieve fractions using
195 bromoform (density 2.88 g/cm³). The bulk heavy mineral fractions were mounted in piperine
196 (Martens, 1932) between a glass slab and a cover. Identification and quantification were
197 carried out under a petrographic microscope by applying the mid-point ribbon and fleet
198 counting methods. At least 200 grains were counted per sample (Mange and Maurer, 1992; table
199 SI3).

200

Zircon geochronology

201 Zircons were extracted from additional 2-3 kg of fresh sandstones crushed with the SelFrag
202 apparatus. Standard Holman-Wilfley shaking tables and magnetic separation methods (Frantz
203 magnetic separator) delivered mineral fractions enriched in zircons. Zircons of grain fraction
204 60-250 µm (non-magnetic) were handpicked under a binocular microscope. Selected grains
205 were mounted in U- and Pb-free resin, polished to expose the internal parts of the minerals, and
206 coated with carbon. Suitable zircon domains and inclusions were identified by
207 cathodoluminescence (CL) and backscattered-electron (BSE) images obtained by split screen
208 on a CamScan CS44 Scandunk Viermalning electron microscope (SEM; Tescan a.s., Brno) at
209 ETH Zurich, and analyzed with laser ablation inductively coupled mass spectrometry (LA-ICP-
210 MS) at ETH Zurich with an excimer laser (ArF 193 nm). Zircon characteristics were interpreted
211 following the approach by Corfu and others (2003).

212 The laser ablation ICP-MS analyses were performed with a RESolution (Australian Scientific
213 Instruments/Applied Spectra) laser ablation system at ETH Zurich equipped with a dual-volume

214 S-155 ablation cell (Laurin Technic, Australia), coupled to an Element XR (Thermo Scientific,
215 Germany) sector-field ICP-MS. A laser repetition rate of 5 Hz, energy density of ca. 2.5 J/cm²,
216 and spot diameter of 30 µm were applied. The carrier gas consisted of high-purity He (ca. 0.7
217 L/min) and make-up Ar (ca. 1 L/min) from the ICP-MS. Data acquisition time per spot was
218 about 1 minute (30 s gas blank + 30 s ablation). The data were processed with the Igor Pro Iolite
219 v2.5 software (Hellstrom and others, 2008), using the VizualAge data reduction scheme (Petrus
220 and Kamber, 2012). Laser-induced element fractionation (after Paton and others, 2010),
221 instrumental mass discrimination, and drift through the analytical session were corrected by
222 standard bracketing against zircon reference material “GJ-1” (using isotope ratios
223 recommended by Horstwood and others, 2016). The quoted uncertainties for each individual
224 analysis correspond to the internal error and propagated uncertainty based on the scatter of the
225 primary reference material (see Paton and others, 2010). The accuracy and reproducibility
226 within each run of analysis were monitored by periodic measurements of zircon reference
227 materials “Plešovice” (337 Ma; Sláma and others 2008) and “91500” (1065 Ma; Wiedenbeck
228 and others 1995). More details about the analytical conditions, the data from unknowns and
229 reference materials are provided in table S5. Concordia diagrams were obtained using
230 ISOPLOT v.3.0 (Ludwig, 2003). A concordant age is given by the overlap of the error ellipse
231 with the Concordia age curve. In this study, ages with discordance >5% are not included for
232 further interpretation (table S4). The frequency U-Pb age distribution diagram or probability
233 density plot described by Ludwig (2003) includes a histogram representing the number of
234 individual zircon grains within a short age range, and the probability curve depicts the mean
235 age peaks of the contained age populations in one sample.

236 *Zircon Lu-Hf isotopic analyses*

237 Lu–Hf analyses were obtained by laser ablation multi-collection inductively coupled plasma
238 mass spectrometry (LA-MC-ICP-MS) at ETH Zurich, using a RESolution (ASI/Applied
239 Spectra) excimer (ArF) 193 nm laser ablation system attached to Nu Plasma II (Nu Instruments)

240 multi-collector ICP-MS. A laser repetition rate of 5 Hz, a spot diameter of 50 μm , and a laser
241 energy density of ca. 4 J/cm^2 was applied. Ablation was performed in a dual-volume, fast-
242 washout S-155 ablation cell (Laurin Technic) fluxed with carrier gas consisting of ca. 0.37
243 L/min He and make-up gas consisting of ca. 0.92 L/min Ar. A typical run consisted of 40 s of
244 background measurement followed by 40 s of sample ablation with an integration time of 0.5
245 s. The ICP-MS instrument was optimized for maximum sensitivity on ^{180}Hf , and optimal peak
246 shape and coincidence for Yb, Lu, and Hf isotopes. The list of acquired masses and
247 corresponding Faraday cups are provided in [table S7](#).

248 The resulting intensities were processed offline with the Igor Pro Iolite v2.5 software, using an
249 in-house data reduction scheme ([Hellstrom and others, 2008](#)). Background-subtracted
250 intensities were used to calculate isotope ratios, which were corrected for instrumental mass
251 bias by normalization to the natural abundance ratios $^{173}\text{Yb}/^{171}\text{Yb}$ (for both Yb and Lu isotope
252 ratios) and $^{179}\text{Hf}/^{177}\text{Hf}$ (for Hf isotope ratios) of [Chu and others \(2002\)](#). The isobaric
253 interferences of ^{176}Yb and ^{176}Lu on ^{176}Hf were subsequently corrected using the natural
254 abundance ratios of $^{176}\text{Yb}/^{173}\text{Yb} = 0.79502$ and $^{176}\text{Lu}/^{175}\text{Lu} = 0.02656$. Accuracy and external
255 reproducibility of the method were controlled by repeated analyses of reference zircon
256 standards “GJ” ([Morel and others, 2008](#)), “Plešovice” ([Sláma and others, 2008](#)), “Mud Tank”
257 and “Temora” ([Woodhead and Hergt, 2005](#)), and “QGNG” ([Woodhead and others, 2004](#)). The
258 $^{176}\text{Hf}/^{177}\text{Hf}$ ratios obtained on all reference materials are within uncertainties identical to the
259 recommended values. Calculations of initial $^{176}\text{Hf}/^{177}\text{Hf}$ ratios and $\epsilon\text{-Hf}$ were performed
260 applying the ^{176}Lu decay constant of [Scherer and others \(2001\)](#) and $^{176}\text{Hf}/^{177}\text{Hf}$ and $^{176}\text{Lu}/^{177}\text{Hf}$
261 values for the chondritic uniform reservoir (CHUR) recommended by [Bouvier and others](#)
262 ([2008](#)). The quoted uncertainties on initial Hf isotopic composition correspond to the analytical
263 uncertainty and the average 2 S.D. ([table S6](#)). Reproducibility of the initial Hf isotopic
264 compositions of all measured zircon reference materials ([table S7](#)), were propagated by

265 quadratic addition. More details about the analytical conditions, the data from unknowns, and
266 reference materials are provided in [table S7](#).

267

268

RESULTS

269

Sandstone frame work

270 The studied Triassic and Jurassic sandstones are mainly classified as feldspathic litharenite and
271 lithic arkose ([fig. 4A](#)). Cretaceous sandstones are classified as subarkose and sublitharenite.

272 The Paleocene sandstone sample is a litharenite. Quartz grains of all samples are mostly
273 monocrystalline (~50% in Triassic, 50-75% in Middle Jurassic, and >80% in Cretaceous and
274 Paleocene sandstones; [figs. 4A, 4B, 4C and table S2](#)). K-feldspars are dominant in feldspar

275 grains (~73% in the Jurassic, and ~80% in the Cretaceous and Paleocene sandstones; [fig. 4B](#)
276 [and table S2](#)). From Jurassic to Cretaceous and Paleocene sandstones, the percentage of

277 monocrystalline quartz and K-feldspar grains increases, while the percentage of polycrystalline
278 quartz and plagioclase grains decreases ([figs. 4C and 4d](#)). The Triassic sandstone is rich in

279 volcanics (~55%), with less sedimentary and metamorphic lithics (~45%). Jurassic sandstones
280 are composed of sedimentary (~60%), volcanic (~30%), and metamorphic (~10%) rock

281 fragments. In Cretaceous and Paleocene sandstones, sedimentary and metamorphic rock
282 fragments (together >95%) are dominant, while volcanic lithics are less common (<5%). In

283 both Jurassic and Cretaceous sandstones, sedimentary lithic grains mainly comprise siltstone,
284 limestone, and dolomite, whereas metamorphic lithic grains generally consist of low-grade to

285 medium-grade slate, phyllites, and schists. Volcanic lithic grains mainly represent andesite and
286 volcanic glass ([figs. 4C, 4D and table S2](#)). Dominantly volcanic lithics ([figs. 4C and 4D](#)) and

287 pyroxene minerals ([fig. 5](#)) point toward a transitional arc as the source terrane of the Triassic
288 sandstone, whereas Jurassic samples point to a dissected arc and mixed nature of the source

289 terranes ([figs. 4E and 4F](#)). Cretaceous and Paleocene sandstones show a recycled orogenic

290 source based on their rock fragment distributions, indicating mineralogical maturity (figs. 4E
291 and 4F).

292

293 *Heavy mineral study*

294 Heavy mineral assemblages of sandstones from the Kopet Dagh are characterized as follows:

295 (1) A group of ultra-stable heavy minerals including zircon, tourmaline, and rutile (ZTR),
296 sphene, monazite, and less stable apatite that are typically derived from acidic to intermediate
297 magmatic rocks and recycled siliciclastic sandstones. The amount of ZTR ranges from 22% of
298 total grain count in the Triassic sandstone and reaches up to 53% of total grain count in Jurassic
299 and up to 74.5% in Cretaceous sandstones (Fig. 5). The percentage of apatite grains ranges
300 between 9% in Triassic and 17.5% in Jurassic sandstones (fig. 5 and table S3). (2) A group of
301 meta-stable heavy minerals including pyroxene and amphibole attributed to mafic magmatic
302 source rocks. The amount of pyroxene in Triassic sandstones reaches up to 64% of the total
303 heavy mineral grains, whereas Jurassic and Cretaceous sandstones contain 24% and 1.5%
304 pyroxene, respectively (fig. 5 and table S3). (3) A third group describes meta-stable minerals,
305 including the epidote group, garnet, chloritoid, andalusite, and staurolite, that are common
306 heavy minerals related to medium- to high-grade metamorphic rocks. The percentage of these
307 minerals varies from 1% of total counted grains in Triassic sandstones to 17% of total counted
308 grains in Paleocene sandstones (fig. 5 and table S3). (4) Fourth, Cr-spinel (up to 4% of total
309 grain count; fig. 5 and table S3) indicates mafic-ultramafic rocks as source of detritus.

310

311 *Detrital zircon U-Pb ages*

312 Laser ablation ICP-MS U-Pb dating on 2478 detrital zircon grains was conducted on one Lower
313 Triassic low-grade meta-sandstone (Qara-Gheitan Formation), five Jurassic turbiditic
314 sandstones (Kashafrud Formation), three Lower Cretaceous fluvial-deltaic sandstones (Shurijeh

315 Formation), and one Paleocene continental sandstone (Pestehleigh Formation; figs. 2, 3, 6, 7, 8,
316 9, and tables S1 and S4).

317 Late Triassic (217 Ma) and early Paleoproterozoic (2374 Ma) zircons are the youngest and
318 oldest zircons of 232 analysed detrital zircons from the Lower Triassic sandstone (fig. 7; sample
319 KD-08). This sandstone reveals a single probability peak at 337 Ma (early Carboniferous). Five
320 Jurassic sandstones show zircon age distributions between 154 Ma (Late Jurassic) and 2966 Ma
321 (Meso-Archean) based on a total of 1191 dated grains (Fig. 7; samples KD-01, KD-03, KD-06,
322 KD-10, KD-15). Detrital zircons of the Jurassic sandstones are peaking at 232 Ma (Early
323 Triassic), 343 Ma (early Carboniferous), 434 Ma (early Silurian), 643 Ma (middle
324 Neoproterozoic), and 1833 Ma (late Paleoproterozoic; Fig. 7). A total of 865 zircons from three
325 Lower Cretaceous sandstones revealed ages between Early Cretaceous (116 Ma) to
326 Mesoarchean (2861 Ma) (fig. 8; samples KD-02, KD-09, KD-18). The Lower Cretaceous
327 sandstones show a similar zircon age distribution as the Jurassic sandstones, but with an
328 additional zircon probability peak at 132 Ma (Early Cretaceous; fig. 8). The Paleocene
329 sandstone shows ages ranging between 119 Ma (Early Cretaceous) and 2781 Ma (Neoproterozoic)
330 from 190 dated zircon grains (fig. 8; sample KD-14). The probability peaks of Paleocene
331 sandstone zircon ages are similar to the ones of Cretaceous sandstones (fig. 8). The combined
332 zircon ages call attention to late Paleoproterozoic, middle Neoproterozoic, early Silurian, early
333 Carboniferous, Late Triassic, and Early Cretaceous protoliths in the potential source areas (figs.
334 9 and 10).

335

336 *Hf isotopic composition*

337 Hafnium isotopic composition analysis was conducted on 151 concordantly dated zircon grains
338 with internal growth structures that were large and thick enough to be examined with laser
339 beams. The overall Hf isotopic compositions (epsilon hafnium values: $\epsilon\text{-Hf}(t)$) of Triassic to
340 Paleocene sandstones range from -23 to +15 (fig. 11). Late Paleoproterozoic zircon grains

341 demonstrate evolved $\epsilon\text{-Hf}_{(t)}$ values from -13 to +1. Early Silurian zircons show a mixed positive
342 and negative pattern of $\epsilon\text{-Hf}_{(t)}$ from -2 to +7 (fig. 11). Early Carboniferous zircons show positive
343 $\epsilon\text{-Hf}_{(t)}$ values of -1 to +12 with four grains between -17 and -4 (fig. 11). Similar to early Silurian
344 zircons, the Late Triassic zircons show mixed positive and negative $\epsilon\text{-Hf}_{(t)}$ values from -3 to +7
345 with one outlier of -15. The only analysed Early Cretaceous zircon shows a $\epsilon\text{-Hf}_{(t)}$ value of +7.6
346 (fig. 11).

347

348

DISCUSSION

349 The presented multi-proxy study of Triassic to Palaeocene sandstones from the Kopet Dagh
350 mountains refine our understanding of the corresponding sediment sources and allows for an
351 improved reconstruction of the tectono-sedimentary evolution of the tectonically inverted
352 Kopet Dagh Basin.

353

Evidence from sandstone frame work

354 The sandstone modal framework analysis of the sandstone lens within a Triassic
355 polymathic conglomerate (Qara-Gheitan Formation) stipulate a transitional
356 magmatic arc provenance (figs. 4E and 4F). However, petrographic analyses on the
357 Qara-Gheitan sandstones by Ghaemi (2009) specify provenance from a dissected
358 magmatic arc. Additionally, geochemical analysis of large granodiorites and
359 tonalities pebbles of the Qara-Gheitan conglomerate by Zanchetta and others, (2013)
360 suggests an active margin setting with prevailing of I-type plutonic and high-K
361 volcanic rocks. The discrepancy between our results and previously published data
362 can be explained by sampling of different sandstone lenses within the
363 inhomogeneous alluvial Qara-Gheitan Formation.

364 The mineralogical composition of Middle Jurassic sandstones of the Kashafrud
365 Formation composed of sedimentary and volcanic lithic fragments with subordinate

366 metamorphic lithics (figs. 4C and 4D) suggest a provenance from a dissected
367 magmatic arc and mixed-recycled orogeny (figs. 4E and 4F). Modal framework and
368 paleocurrent analysis along with stratigraphic studies of Middle Jurassic sandstones
369 by Poursoltani and others, (2007) and Poursoltani and Gibling (2011) are in
370 agreement with our results. In addition, they suggested that the detritus was supplied
371 from an active margin in the southeast to the extensional Kopet Dagh Basin in the
372 northwest through channel systems to slope and basin-floor fans forming fluvial
373 (basal conglomerate), deltaic (sandstone), and proximal to distal turbiditic
374 (sandstone and mudstone) facies.

375 Lower Cretaceous sandstones (Shurijeh Formation) show higher mineralogical and
376 textural maturity compared to Middle Jurassic sandstones (figs. 4A, 4B, 4C and 4D)
377 and indicate a recycled orogenic provenance (figs. 4E and 4F). In addition, modal
378 framework analysis by Mortazavi and others (2013) on Lower Cretaceous
379 sandstones supports a provenance characterized by recycled orogenic belts and
380 craton interior signature. Paleocurrent directions of Lower Cretaceous sandstones
381 suggest that sediments were supplied from the south and southwest into the Kopet
382 Dagh Basin (Moussavi-Harami and Brenner, 1990). Therefore, Jurassic sandstones
383 represent a potential source for second-cycle Lower Cretaceous sandstones.

384 Paleocene continental sandstones (Pestehleigh Formation) show relatively higher
385 mineralogical maturity compared to Triassic-Jurassic sandstones (figs. 4A, 4B, 4C
386 and 4D), indicating a recycled orogenic provenance (figs. 4E and 4F). This
387 interpretation agrees with findings of Moussavi-Harami (1993), who suggested that
388 northward and northwestward thinning of coarse-grained Paleocene sandstone layers
389 indicates that the detritus was supplied from the recycling of older sediments in the
390 south and southeast.

Evidence from heavy minerals assemblage

391
392 The heavy mineral assemblage of the Triassic sandstone with dominant pyroxene minerals (fig.
393 5 and table 3) indicates a source from mafic magmatic rocks (probably a magmatic arc), which
394 is in concordance with sandstone modal framework results (figs. 4E and 4F). In addition, the
395 internal textures (CL images) and Th/U ratios of detrital zircons from pebbles of the Triassic
396 conglomerate suggest a magmatic origin for these zircons (Zanchetta and others, 2013). An
397 increase in ultra-stable minerals (ZTR) along with a decreasing percentage of pyroxene
398 minerals in the heavy mineral fraction of Jurassic sandstones (fig. 5 and table 3) imply first-
399 cycle sandstones that were transported over a long distance, or second-cycle sandstone formed
400 by erosion and deposition of pyroclastic-volcanoclastic rocks in a deep-marine basin
401 (Poursoltani and others, 2007). A significant increase of clastic sedimentary lithics compared
402 to the Triassic sandstone and first appearance of garnets in Jurassic sandstones (fig. 5 and table
403 3) point toward metamorphic rocks as a minor additional source of detritus. Poursoltani and
404 Gibling (2011) reported ultra-stable heavy minerals (ZTR), along with apatite and garnet as the
405 main heavy mineral spectrum in the Jurassic sandstones, coinciding with our findings. Early
406 Cretaceous sandstones indicate a high percentage of ultra-stable minerals (ZTR) and a reduced
407 quantity of pyroxene minerals in the heavy mineral assemblages with a much significant
408 intensity. Vice versa, the amount of pyroxene (20%) in Jurassic sandstones decreases to 1.5%
409 in Cretaceous sandstones (fig. 5 and table 3). This indicates that the heavy mineral content was
410 partly controlled by cannibalism and recycling of older sandstones. An increase of
411 monocrystalline quartz content along with a decrease of less stable volcanic lithics in Early
412 Cretaceous sandstones compared to older lithologies provides further evidence for recycling of
413 Jurassic sandstones (table 3). In addition, based on geochemical analysis of Early Cretaceous
414 sandstones, Mortazavi and others, (2013) suggested moderate to high weathering of the
415 sediments during transportation before deposition, which is in agreement with recycling of
416 older sediments. However, the effect of diagenetic processes on the decrease of metastable

417 heavy minerals (mostly pyroxene group) and volcanic lithics, and partial alteration of feldspars
418 to clay minerals should not be ignored (Mortazavi and others, 2014; Moussavi-Harami and
419 Brenner, 1993).

420 Paleocene sandstones indicate a heavy mineral spectrum similar to the one of Early Cretaceous
421 sandstones. However, the amount of garnet increases from 3.5% in Early Cretaceous sandstones
422 to 15.5% in Paleocene sandstones. This implies that erosion of garnet-bearing metamorphic
423 rocks, along with recycling of the older sandstones, were the main sources of detritus in the
424 Palaeocene. The most probable source of garnet and metamorphic lithic grains are Carbo-
425 Permian metamorphic rocks to the southeast of the Kopet Dagh Basin (figs. 2 and 3).

426

427 *Evidence from Detrital Zircon U-Pb Ages and Hf Isotopes*

428 The 2477 analyzed detrital zircon grains from the Lower Triassic, Middle Jurassic, Lower
429 Cretaceous, and Paleocene turbiditic, deltaic-fluvial, and continental sandstones yielded
430 apparent ages ranging from ca. 119 Ma to ca. 2966 Ma (figs. 7, 8, 9 and 10). Six major U-Pb
431 age groups are identified in the Kopet Dagh detrital zircon.

432 *Mesoarchean to Paleoproterozoic Zircons*

433 The oldest zircons create a broad heterogeneous age spectrum from Mesoarchean to
434 Paleoproterozoic (2966-1600 Ma) with a main peak in the Paleoproterozoic (1833 Ma; fig. 9).
435 Zircons of this population contain ~11% of total dated zircons and are characterized by anhedral
436 to rounded grains without internal zoning (>80%), and a wide range of Th/U ratios (between
437 0.01 and 2.39; fig. 6 and table S4). The $\epsilon\text{-Hf}(t)$ values of Mesoarchean to early Paleoproterozoic
438 zircons represent a mixture of positive and negative values (-6.9 to +10.3), while late
439 Paleoproterozoic zircons show dominantly negative values (-13.6 to -2.3) with only one grain
440 of positive value (+ 1.4; fig. 11 and table S6). Possibly, these are inherited magmatic zircons or
441 fragments of older reworked zircons from crystalline rocks of a craton. Because of the
442 probability of repetitive reworking of detrital zircons through several tectonic cycles since the

443 Archean, the possible metamorphic/igneous source area remains undefined. However, similar
444 age spectra (with a major peak at 2.5 Ga and a minor peak at 1.8 Ga) were reported in
445 Neoproterozoic to Cambrian, lower Permian and Middle Jurassic, and Devonian sandstones of
446 the Alborz mountains (Horton and others, 2008; Moghadam and others, 2017). Also,
447 Mesoarchean to Paleoproterozoic zircons were reported in Cretaceous to Eocene sandstones of
448 Azerbaijan Basin (Mohammadi and others, 2020a), Eocene to Oligocene sandstones from the
449 Sistan Basin (Mohammadi and others, 2016a), and Cretaceous to Miocene sandstones of the
450 Makran Basin in Iran (Mohammadi and others, 2016b; 2017). In addition, similar age spectra
451 were reported in Paleozoic to Triassic sandstones of the Taurides mountains in Turkey (Abbo
452 and others, 2015). There are no reported occurrences of Archean to Paleoproterozoic zircons in
453 the Iranian or Anatolian continental crust (for example Ramezani and Tucker, 2003;
454 Moghadam, and others, 2017; Moghadam, and others, 2020). However, Archean to
455 Paleoproterozoic zircons with main peaks at 2.5 Ga and 1.8 Ga were identified from
456 metamorphic basement of the Afghan block (Faryad and others, 2016), the Tibetan terranes
457 including Songpan Ganzi (Weislogel and others, 2006) and Nan Shan-Qilian Shan-Altun Shan
458 (Gehrels and others, 2011), and the Tarim craton (Gehrels and others, 2011). The lack of a 1.8
459 Ga and presence of a 1.9 Ga zircon age peak in the Archean to Paleoproterozoic age spectrum
460 of Iberia and African Neoproterozoic to Cambrian sandstones (Abdelsalam and others, 2002)
461 excludes the Saharan meta-craton as a potential source area. In fact, the source of Archean to
462 Paleoproterozoic zircons in the Kopet Dagh strata remains enigmatic. Restricting conjecture to
463 the closest cratons points to the nearby Afghan Block with Archean to Paleoproterozoic
464 metamorphic and igneous rocks (Faryad and others, 2016) as an original source area, which
465 supplied detritus to the Kopet Dagh Basin through repetitive recycling of sedimentary
466 successions (fig. 10).

467

Neoproterozoic Zircons

468 The second age population includes Neoproterozoic detrital zircons (1000-542 Ma) with a main
469 peak at 643 Ma (fig. 9). This population represents ~18% of the total dated zircons.
470 Neoproterozoic zircons are mostly rounded or anhedral grains with sporadic magmatic or
471 metamorphic growth zoning or lack zoning (fig. 6). Similar to Archean to Paleoproterozoic
472 zircons, Neoproterozoic zircons show Th/U ratios with a wide range between 0.01 and 2.92
473 (table S4), which indicate both magmatic and metamorphic sources. Due to different grain
474 shapes and internal growth structures, Neoproterozoic zircons are probably remnants of older
475 recycled grains or inherited cores from continental crystalline rocks. The $\varepsilon\text{-Hf}(t)$ values of
476 Neoproterozoic zircons range between -19.3 and +11.3 (fig. 11 and table S6). Neoproterozoic
477 (Pan-African) detrital zircons were reported from several Gondwana-related tectono-
478 stratigraphic domains such as Iran (for example Hassanzadeh and others, 2008; Hosseini and
479 others, 2015; Nutman and others, 2014; Ramezani and Tucker, 2003), the Taurides (Abbo and
480 others, 2015), Libya (Meinhold and others, 2013), the Arabian-Nubian Shield (Morag and
481 others, 2012), and north Africa (Abati and others, 2010; Avigad and others, 2012).
482 Neoproterozoic detrital zircons with a main peak at 559-537 Ma (Cadomian age) were
483 furthermore reported from Alborz and Zagros Neoproterozoic to Paleozoic strata (Horton and
484 others, 2008; Moghadam and others, 2017). In addition, the Neoproterozoic detrital zircons
485 without internal zoning with a main peak at 640-610 Ma were reported from Central Iranian
486 Neoproterozoic to Cambrian sandstones (Honarmand and others, 2016). Nutman and others
487 (2014) reported 637 Ma U-Pb ages of magmatic zircons from the Central Iran crystalline
488 basement (Chadegan orthogneiss). Therefore, repetitive recycling of the Cryogenian to
489 Ediacaran crystalline basement of Central Iran is a plausible source of the Neoproterozoic
490 detrital zircons of Kopet Dagh strata (fig. 10).

491

492
493 The third age group denotes Ordovician to Silurian zircons (465-422 Ma) with a peak at 437
494 Ma with subhedral to euhedral shapes, dominantly magmatic zoning, and Th/U ratios above 0.5
495 (fig. 6 and table S4), demonstrating an abundant magmatic source. This population contains
496 ~9.3% of the total dated zircons (table S4). $\epsilon\text{-Hf}(t)$ values (-1.9 to +6.7, only one grain with -
497 8.9; fig. 11 and table S6) suggest rift-related magmatic rocks as detritus source (for example
498 Vervoort and Blichert-Toft, 1999). Presented zircon ages correspond to major rifting episodes
499 that occurred from late Ordovician to Silurian along the northern Gondwana margins
500 (Moghadam and others, 2020; Ruban, 2007; Stampfli, 2000), which correspond to the
501 separation of the Hun terranes from Gondwana along with the opening of Paleo-Tethys Ocean
502 (Moghadam and others, 2020). Late Ordovician to middle Silurian rift-related magmatism was
503 furthermore reported from northern Central Iran (for example Nowrouzi and others, 2014),
504 eastern Alborz (Derakhshi and Ghasemi, 2015; Moghadam and others, 2020), and Anatolia (for
505 example Topuz and others, 2020; Topuz and others, 2021). In the eastern Alborz, basaltic and
506 basaltic-andesitic magmatism during the Silurian were reported (local name: Soltan Maidan
507 Complex; Derakhshi and Ghasemi, 2015; Ghavidel-Syooki and others, 2011). Geochemistry of
508 this volcanic units shows alkaline affinity with an intra-plate and continental-rift setting
509 (Derakhshi and Ghasemi, 2015). The zircon U-Pb ages of rounded granitic pebbles from
510 conglomerate layers at the base and in middle of the afore-mentioned volcanic unit indicate an
511 early Silurian genesis (434 ± 6.4 Ma; Ghavidel-Syooki and others, 2011). Furthermore, zircon
512 U-Pb ages of A-type granitic pebbles from Devonian sandstones (Padeha Formation) in the
513 eastern Alborz indicate early Silurian ages (441 ± 3.1 Ma) with $\epsilon\text{Hf}(t)$ values ranging from +8.2
514 to +2.5 (Moghadam and others, 2017). The zircon U-Pb age of granitic pebbles from both
515 Silurian and Devonian successions fit well with the main peak (437 Ma) of the late Ordovician
516 to Silurian detrital zircon population of the Kopet Dagh sandstones (fig. 9). Furthermore,
517 Ordovician to Silurian detrital zircons with a main peak at 435 Ma and $\epsilon\text{Hf}(t)$ values of -3.6 to

518 +6.3 from late Ordovician (Qelli Formation) sandstones from eastern Alborz were reported by
519 [Moghadam and others \(2017\)](#). Taking into consideration all above-mentioned criteria as well
520 as a short transport distance between eastern Alborz and Kopet Dagh, Ordovician to Silurian
521 detritus of the Kopet Dagh Basin was most probably supplied through recycling of Ordovician
522 to Devonian sandstones from the eastern Alborz.

523

524 *Devonian to Carboniferous Zircons*

525 The fourth age population (385-305 Ma) with a main peak at 340 Ma ([fig. 9](#)) exhibits dominantly
526 euhedral zircon grains (~75%) with magmatic internal texture and high Th/U ratios (> 0.5; [fig.](#)
527 [6 and table S4](#)). This age spectrum comprises ~17% of the total analysed detrital zircons ([table](#)
528 [S4](#)). Devonian to Carboniferous zircons display general $\varepsilon\text{-Hf}_{(t)}$ values between -1.4 and +15.3
529 and three grains between -12.1 and -17.7 ([fig. 11 and table S6](#)). Dominant positive $\varepsilon\text{-Hf}_{(t)}$ values
530 suggest magmatic zircons with a depleted mantle signature as observed in mafic-ultramafic
531 oceanic crustal and ophiolitic rocks ([Patchett, 1983](#); [Blichert-Toft and Albarède, 1997](#)).
532 Carboniferous detrital zircons with a main peak at 344-347 Ma from Cretaceous turbiditic
533 sandstones of Azerbaijan (NW Iran) were reported by [Mohammadi and others \(2020a\)](#).
534 However, a further potential source area includes the northeast of Iran, where Devonian to
535 Carboniferous magmatic rocks were reported from the Aghdarband (Darrehanjir) ophiolite
536 complex ([Moghadam and others, 2015a](#); [Zanchetta and others, 2013](#)). The Aghdarband window
537 in the southeast of the Kopet Dagh Basin contains Permo-Triassic undifferentiated sedimentary,
538 low-grade metamorphic, and igneous rocks. Igneous rocks include diabase, gabbro, ultramafic,
539 and volcanoclastic sediments ([figs. 2 and 3](#); [Afshar-Harb 1982a](#); [Behrouzi and others, 1993](#)).
540 In the southern part of the Aghdarband complex, an upper Permian to Lower Triassic polymictic
541 conglomerate (Qara-Gheitan Formation; [Behrouzi and others, 1993](#)) overthrusts the Triassic
542 unit. The composition and age of the Aghdarband basement is generally poorly constrained.
543 However, the exposed Devonian to Carboniferous meta-sediments in the north of the

544 Aghdarband complex (Behrouzi and others, 1993; Zanchetta and others, 2013) indirectly
545 indicates the basement age of this complex. Besides this, zircon U-Pb ages of plagio-granitic
546 dikes and ferro-diorites (382-380 Ma) from the Darrehanjir ophiolites also indicate a Late
547 Devonian age for this basement complex (Moghadam and others, 2015a). Furthermore, the
548 zircon U-Pb ages of granitoid cobbles in the Qara-Gheitan conglomerate indicate a late
549 Carboniferous (314 Ma) age (Zanchetta and others, 2013). The wide range of Devonian to
550 Carboniferous rocks in the Aghdarband complex is in concordance with detrital zircon ages in
551 the Kopet Dagh sandstones (fig. 9). Supportively, positive zircon $\epsilon\text{Hf}(t)$ values (+13.8 and
552 +14.9) of dated plagio-granites and ferro-diorites suggest a depleted mantel signature
553 (Moghadam and others, 2015a). Devonian and Carboniferous granites and granitoids were
554 reported from the Karakum-Mangyshlak and Karabogaz micro-blocks in Central Asia (for
555 example Natalin and Şengör, 2005). Widespread late Paleozoic to early Mesozoic volcanism in
556 the Turan Block reported by Natalin and Şengör (2005) could represent another possible source
557 for early Carboniferous detrital zircons from the Kopet Dagh. However, the calc-alkaline
558 affinity of these rocks indicates a magmatic arc setting, which is not in agreement with the Hf
559 isotopic composition of detrital zircons. In addition, the reported general SE-NW paleocurrent
560 direction in the Kopet Dagh sandstones (Poursoltani and others, 2007) strengthen the case that
561 the Devonian to Carboniferous detritus was supplied from the Aghdarband Complex rather than
562 from the Turan Block.

563

564 *Permian to Triassic Zircons*

565 The fifth zircon age spectrum in the Kopet Dagh sandstones (300-200 Ma) with a peak at 233
566 Ma (fig. 9) covers ~25.4% of the total dated zircons (table S4). Zircon grains of this population
567 exhibit euhedral shapes (>80%) with magmatic zoning (fig. 6). Furthermore, over 90% of
568 zircons in this spectrum indicate a high Th/U ratio (>0.55; table S4), which denotes magmatic
569 origin. $\epsilon\text{-Hf}(t)$ values of Permo-Triassic zircons range between -5.9 and +10.3 (>80%) with two

570 grains with values of -15 and -15.2 (fig. 11 and table S6). Similar to Devonian to Carboniferous
571 zircons, dominant positive $\varepsilon\text{-Hf}(t)$ values of the Permo-Triassic zircons propose a depleted
572 mantle signature related to mafic-ultramafic rocks from oceanic crust and ophiolite complexes
573 (Patchett, 1983; Blichert-Toft and Albarède, 1997). Both Permian and Triassic ages were
574 reported from the Fariman and Mashhad complexes in the south of the Kopet Dagh Basin (for
575 example Chiu and others, 2017; Ghazi and others, 2001; Karimpour and others, 2010; Kozur
576 and Mostler, 1991; Mirnejad and others, 2013; Moghadam and others, 2015a; Topuz and
577 others, 2018; Zanchetta and others, 2013).

578 The Fariman complex consists of two units: i) The lower metamorphosed unit composed of
579 green schists, metabasites, mafic-ultramafic rocks, and serpentinites, and ii) the upper volcanic-
580 sedimentary unit. Conodonts and radiolarian fossils reveal a late early Permian to middle
581 Permian age for the Fariman complex (Zanchetta and others, 2013). In addition, $^{40}\text{Ar}\text{-}^{39}\text{Ar}$
582 dating of magmatic hornblende from olivine cumulates yielded early Permian (276 ± 4 Ma) ages
583 for the Fariman complex (Topuz and others, 2018). Furthermore, andesitic dykes dated at 211
584 Ma indicate Late Triassic magmatic activity in the Fariman complex (Zanchetta and others,
585 2013). The Fariman complex has been interpreted as a continental magmatic arc with an intra-
586 arc basin (Zanchetta and others, 2013), a subduction- and plume-related Paleo-Tethyan
587 ophiolite (Moghadam and others, 2015a), as well as a fragment of an oceanic plateau (Topuz
588 and others, 2018).

589 The Mashhad Complex in the south of the Kopet Dagh Basin represents a remnant
590 of the Paleo-Tethys oceanic crust and includes three members: i) The lower member
591 with ultramafics, gabbro-diorites, pillow lavas, marbles, and meta-sediments
592 (Moghadam and others, 2015a). An early Permian age was suggest for the Mashhad
593 ophiolite based on the study of pelagic conodonts (Kozur and Mostler, 1991) and
594 $^{40}\text{Ar}\text{-}^{39}\text{Ar}$ ages of hornblende from gabbros (287.6 Ma and 281.7; Ghazi and others,
595 2001). ii) The middle member consists of Permo-Triassic deep-water clastic

596 turbidites and metamorphosed sedimentary rocks (accretionary prism). iii) The
597 upper member exhibits arc-related acidic to intermediate pyroclastic rocks
598 (Moghadam and others, 2015a). Post-collisional intrusive bodies crosscut the
599 Mashhad complex that contain several granitoid and monzogranite suites with zircon
600 U-Pb ages between 200 to 217 Ma (Karimpour and others, 2010; Mirnejad and
601 others, 2013). In addition, the Mashhad granodiorite and Mashhad diorite were dated
602 to 203 Ma and 219 Ma, respectively, with $\varepsilon\text{-Hf}_{(t)}$ values between -3 to +3 (Chiu and
603 others, 2017). The complex is unconformably is covered by Jurassic terrigenous
604 sediments of the Kashafrud Formation. The southeastward continuation of the
605 Mashhad Complex, which is known as the Torbat-e-Jam Complex, was also
606 intruded by post-collisional I-type granitites (217 Ma; Zanchetta and others, 2013)
607 and gabbro-quartz diorite (215 Ma; Ghavi and others, 2018) of similar ages.
608 According to Zanchetta and others (2013), the age range of post-collisional
609 intrusions indicates that the obduction of the Mashhad ophiolites occurred before the
610 Late Triassic (Norian-Rhaetian).

611 The age spectrum and dominant positive $\varepsilon\text{-Hf}_{(t)}$ values of Permian to Triassic
612 zircons, together with the presence of chromium spinel in the heavy mineral
613 assemblage, indicates an exposure of mafic-ultramafic rocks in the source area. This
614 interpretation is in agreement with unroofing and consequent deposition of the
615 Mashhad and Fariman Complexes, and a remnant of Triassic continental arc in the
616 Kopet Dagh Basin. Northward directed paleocurrents and the presence of Norian
617 granodiorite pebbles (211 Ma) in the conglomerate underlying Jurassic sandstones
618 indicate erosion and unroofing of post-collisional intrusive bodies and ophiolitic
619 host rocks in Middle Jurassic time (for example Zanchetta and others, 2013; Zanchi
620 and others, 2009; Zanchi and others, 2016).

621

Early Cretaceous Zircons

622
623 The sixth and youngest age population (146-119 Ma) with a main peak at 132 Ma (fig. 9)
624 contains only 1.5% of the total dated zircon grains (table S4). Zircons grains of this population
625 are euhedral in shape, exhibit magmatic zoning, and show Th/U ratios >0.3 (fig. 6 and table
626 S4). $\epsilon\text{-Hf}(t)$ values (+7.6) of this population indicate non-depleted mantle signatures (fig. 11 and
627 table S6). Occurrence of Early Cretaceous zircons in Lower Cretaceous sandstones (samples
628 KD-09 and KD-18) and the Paleocene sandstone (sample KD-14; figs. 8 and 10) indicate syn-
629 sedimentary magmatic activity in the source area. Furthermore, the presence of detrital zircons
630 with early Aptian ages (~124 Ma; average of 8 youngest ages) in Lower Cretaceous sandstones
631 of the Shurijeh Formation indicate maximum depositional ages younger than the previously
632 reported Neocomian stratigraphic age (for example Afshar-Harb 1982a, 1982b; Afshar-Harb
633 and others, 1986; Behrouzi and others, 1993; Mortazavi and others, 2014; Moussavi-Harami
634 and Brenner, 1993). The potential source area of Early Cretaceous zircons could be the Cheste
635 Sharif pluton in north Afghanistan. The Cheste Sharif pluton is located in the Band-e-Bayan
636 Zone (a narrow W-E trending crustal block bounded by the large-scale Hari Rod fault) and is
637 dated at 138 Ma by K-Ar biotite (Debon and others 1987). The presence of Early Cretaceous
638 zircons in Lower Cretaceous sandstones in the eastern Kopt Dagh (fig. 2) is in agreement with
639 a detritus supply from the Band-e-Bayan Zone during Early Cretaceous time. Most probably,
640 Early Cretaceous zircons in the Paleocene sandstones were recycled from Lower Cretaceous
641 sandstones (fig. 10).

642

643 REGIONAL TECTONIC IMPLICATIONS

644 Integrated provenance analysis including zircon ages and isotopic compositions of the
645 Mesozoic to Paleocene detrital material in the Kopet Dagh Basin recorded three main tectonic
646 events in the Phanerozoic (figs. 9 and 11). These events took place in late Ordovician to early
647 Silurian (main peak at 437 Ma), Late Devonian to early Carboniferous (main peak at 340 Ma),

648 and late Permo-Triassic (main peak at 233 Ma). The first event (late Ordovician to early
649 Silurian) corresponds to the separation of Eurasia from Gondwana and the consequent opening
650 of the Paleo-Tethys Ocean (for example Stampfli and Borel, 2002; Stampfli and others, 2013)
651 (fig. 12; 440 Ma). This rifting event is marked by mafic magmatism along the northern margin
652 of Gondwana (Alborz Mountains; Derakhshi and Ghasemi, 2015). Hf-isotopic composition of
653 Kopet Dagh Ordovician-Early Silurian detrital zircons is in agreement with magmatism in a
654 rifting setting.

655 The second recorded event in late Devonian to early Carboniferous coincides with a plutonic
656 episode with A-type (intra-continental) magmatism in NW Iran (Azerbaijan area) and the
657 Sanandaj-Sirjan Zone (for example Mohammadi and others, 2020b). In addition, late Devonian
658 to early Carboniferous ages are represented by a series of mafic and ultramafic rocks (ophiolites
659 and oceanic igneous complexes) from Turkey in the west to Tibet to the east (for example
660 Moghadam and others, 2015b). This event records the first stages of Paleo-Tethys subduction
661 beneath the Turan Block that initiated in Late Devonian (~380 Ma) and continued until Late
662 Triassic (Moghadam and others, 2015a) (fig. 12; 380 Ma). According to Zanchetta and others
663 (2013), the exposed Devonian to Carboniferous Aghdarband and Permo-Triassic Fariman
664 complexes in NE Iran formed as a back-arc and an intra-arc basin, respectively, along the
665 southern margin of the Turan Block, and related volcanoclastic rocks were supplied from an
666 active magmatic arc in the south of the Kopet Dagh Basin (fig. 12; 340 Ma). Contrariwise,
667 Moghadam and others (2015a) argued that the Aghdarband and Mashhad ophiolites in NE Iran
668 were part of the Paleo-Tethys oceanic crust, where related volcanoclastic rocks were supplied
669 from an active magmatic arc (Silk Road Arc) in the north of the Kopet Dagh Basin. New zircon
670 data presented in this study with depleted mantle composition and a main peak at 340 Ma
671 support the hypothesis of Zanchetta and others (2013) and point toward intensified rifting and
672 spreading of a back-arc basin ~40 Myr after the initiation of the Paleo-Tethys subduction (fig.
673 12; 380 Ma).

674 The third event (late Permian to late Triassic) coincides with a continental magmatic arc and
675 the final collision of the Paleo-Tethys (fig. 12; 230 Ma). Rifting of the Iran block from
676 Gondwana in Late Paleozoic forced to northward subduction of the Paleo-Tethys Ocean
677 beneath the Turan (Eurasia) plate, which resulted in the formation of a Permo-Triassic
678 continental magmatic arc. The Mashhad Ophiolite Complex was abducted in early Late Triassic
679 (Carnian-Norian; Moghadam and others, 2015b). Finally, collision between the Iran and Turan
680 plates occurred in Late Triassic (Norian-Rhaetian) and was accompanied by the intrusion of
681 several post collisional plutonic bodies into the magmatic arc and the Mashhad Ophiolite
682 Complex (fig. 12; 230 Ma). Finally, unroofing of the magmatic arc, the Mashhad Ophiolite
683 Complex, and related post-collisional intrusions in Middle Jurassic led to deposition of Middle
684 Jurassic turbiditic sandstones in the Kopet Dagh extensional basin (fig. 12; 165 Ma).

685

686

CONCLUSIONS

687 The present study assesses the provenance of detrital sediments in Triassic to Paleocene
688 turbiditic, fluvial-deltaic and continental sandstones of the Kopet Dagh Basin in northeast Iran.
689 Over 2470 detrital zircons U-Pb ages yield a spectrum of variable protoliths from Archean to
690 Early Cretaceous (2966 Ma – 119 Ma) with main peaks in Paleoproterozoic (1883 Ma),
691 Neoproterozoic (643 Ma), early Silurian (437 Ma), early Carboniferous (340 Ma), Late Triassic
692 (233 Ma), and a minor peak in Early Cretaceous (132 Ma). The combined U-Pb ages and Hf
693 isotope data indicate that protolith rocks belonged to an early Silurian intracontinental rift
694 (opening of Paleo-Tethys), an early Carboniferous rifting of a back-arc (Aghdarband Basin),
695 and a Late Triassic collisional to post-collisional (Paleo-Tethys collision) magmatism. Our
696 results highlight the timing of Aghdarband Basin opening 40 Ma after subduction initiation of
697 the Paleo-Tethys. Cr-spinel and heavy mineral assemblages imply additional detritus sourcing
698 from ophiolites (Mashhad and Fariman complexes). Our analytical results furthermore suggest
699 that the majority of Jurassic sedimentary detritus sourced from protoliths in the south of the

700 Kopet Dagh Basin and deposited in a submarine fan-delta complex. Early Cretaceous detritus
701 was supplied from similar protoliths and furthermore from recycled Jurassic turbiditic
702 sandstones. These sediments were deposited in a fluvial-deltaic complex. The Paleocene
703 continental sediments were mainly supplied from recycling of exposed Jurassic to Early
704 Cretaceous sediments.

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708

709 Supplementary Table Captions

710 Table S1: Geographical location (degree, minute, second), lithology, local formation name, sedimentary
711 environment, age and characteristics of the analyzed samples of eastern Kopet Dagh sandstones.
712 Analysis key: MFA: modal framework analysis, HM: heavy minerals, DZA: detrital zircon age and
713 Hafnium isotopes.

714

715 Table S2: Sandstone framework data and calculated percentages for eastern Kopet Dagh sandstones.

716

717 Table S3: Heavy mineral data, calculated percentages and heavy mineral indexes of eastern Kopet Dagh
718 sandstones.

719

720 Table S4: Zircon U-Pb isotopic data and calculated ages for eastern Kopet Dagh sandstones. C= Coarse
721 grain (>200 μ) and M= Medium grain (100-200 μ).

722

723 Table S5: Zircon U-Pb dating details including sample preparation, analytical conditions, laser ablation
724 system, ICP-MS instrument, data from unknowns and reference materials, and data processing.

725

726 Table S6: Lu-Hf isotopic data and calculated ratios for eastern Kopet Dagh sandstones.

727 Table S7: Zircon Lu-Hf isotopic analyses details including sample preparation, analytical conditions,
728 laser ablation system, ICP-MS instrument, data from unknowns and reference materials, and data
729 processing.

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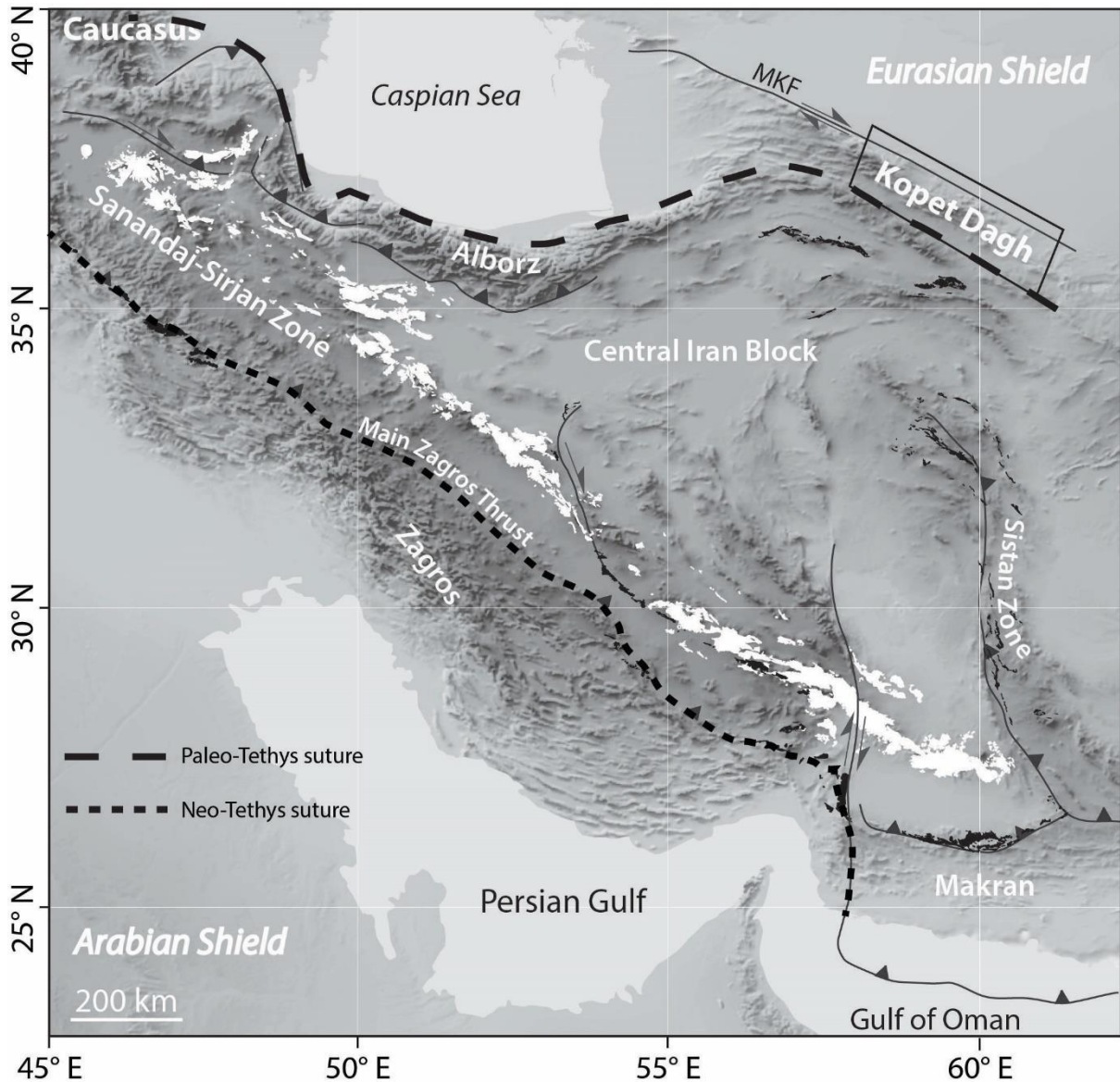
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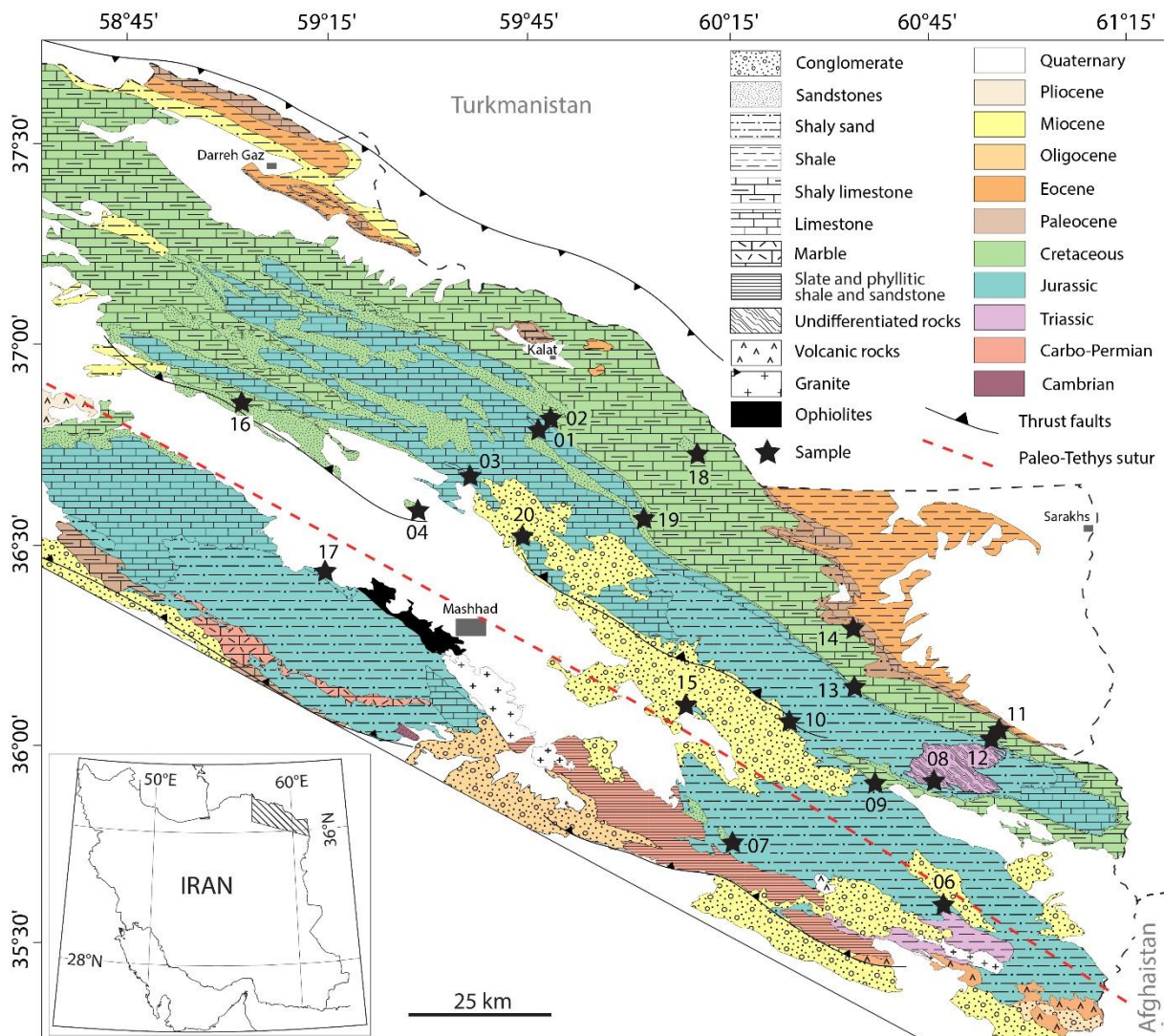
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Fig. 1. Simplified tectonic setting of Iran including Urumieh Dokhtar Magmatic Arc (white) and the Tethys ophiolite complexes (black). Framed: eastern Kopet Dagh Basin. Background: shaded relief map ETOPO1 (<http://www.ngdc.noaa.gov/mgg/global/relief/ETOPO1>).



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1092 Fig. 2. Simplified geologic map of the East Kopet Dagh Basin and samples location.
1093 Stratigraphic ages correspond to the 1:250,000 geological maps of Sarakhs ([Afshar-Harb,](#)
1094 [1982a](#)), Darreh Gaz ([Afshar-Harb, 1982b](#)), Mashhad ([Afshar-Harb and others, 1986](#)), and
1095 Torbat-e-Jam ([Behrouzi and others, 1993](#)).

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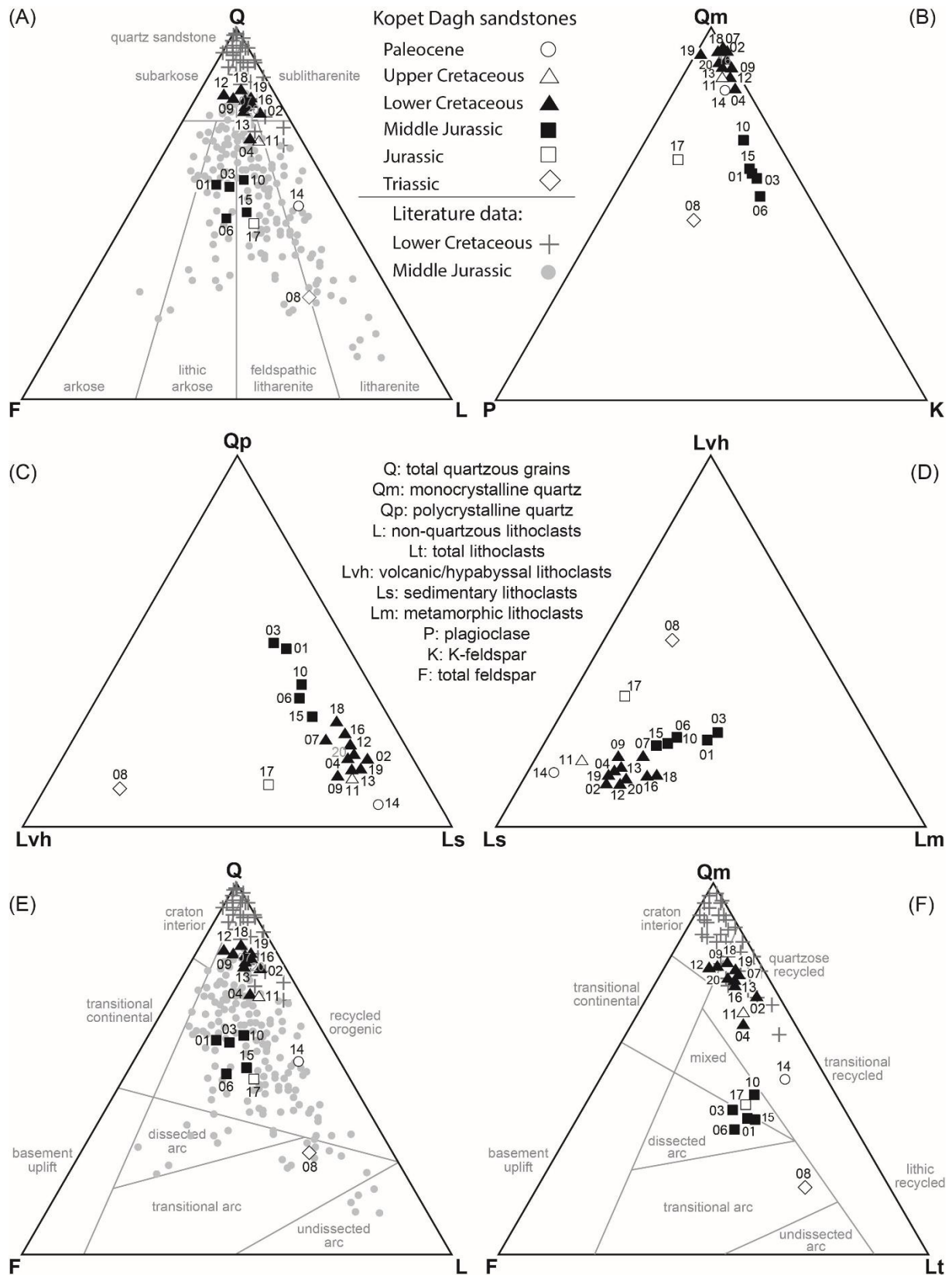
1102 Fig. 3. Composite stratigraphic section of the East Kopet Dagh Basin with local formation

1103 names, sampled formations, and comments on lithology and sedimentary environment.

1104 Stratigraphic ages correspond to the 1:250,000 geological maps of Sarakhs (Afshar-Harb,

1105 1982a), Darreh Gaz (Afshar-Harb, 1982b), Mashhad (Afshar-Harb and others, 1986), and

1106 Torbat-e-Jam (Behrouzi and others, 1993). Black stars: sampled sedimentary sequences.



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Fig. 4. Detrital composition of Kopet Dagh sandstones in ternary classification schemes and

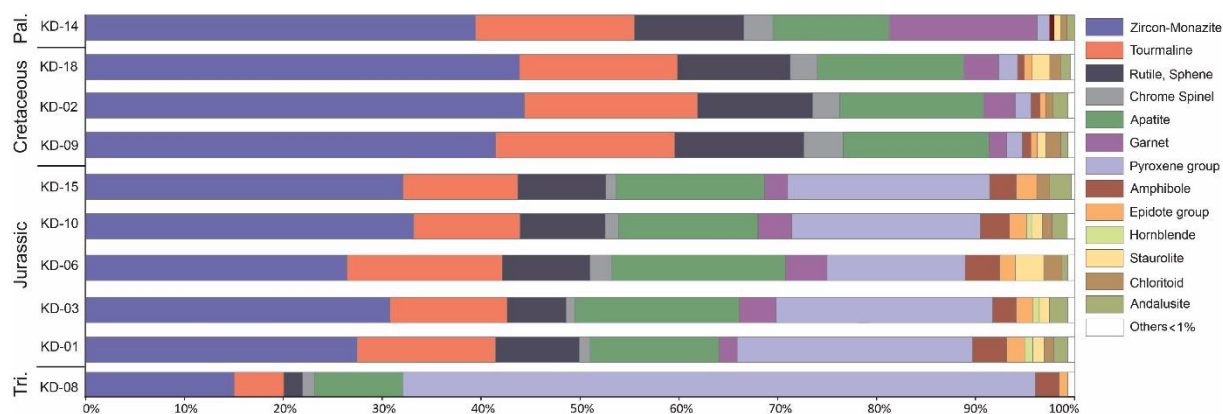
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provenance discrimination diagrams. (A) Q-F-L sandstone classification diagram (Folk, 1980);

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(B) Qm-P-K diagram (after Dickinson and Suczek, 1979); (C) Qp-Lvh-Lsm diagram (after

1111 Dickinson and Suczek, 1979); (D) Lvh-Ls-Lm diagram (after Dickinson, 1985); (E) QFL
 1112 (Dickinson, 1985); and (F) QmFLt (Dickinson, 1985). Literature data of Middle Jurassic and
 1113 Lower Cretaceous sandstones from Poursoltani and Gibling (2011) and Mortazavi and others,
 1114 (2013).
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 1118 Fig. 5. Heavy mineral assemblages of the Kopet Dagh sandstones arranged by stratigraphic ages
 1119 displaying the relative abundance of heavy mineral groups.
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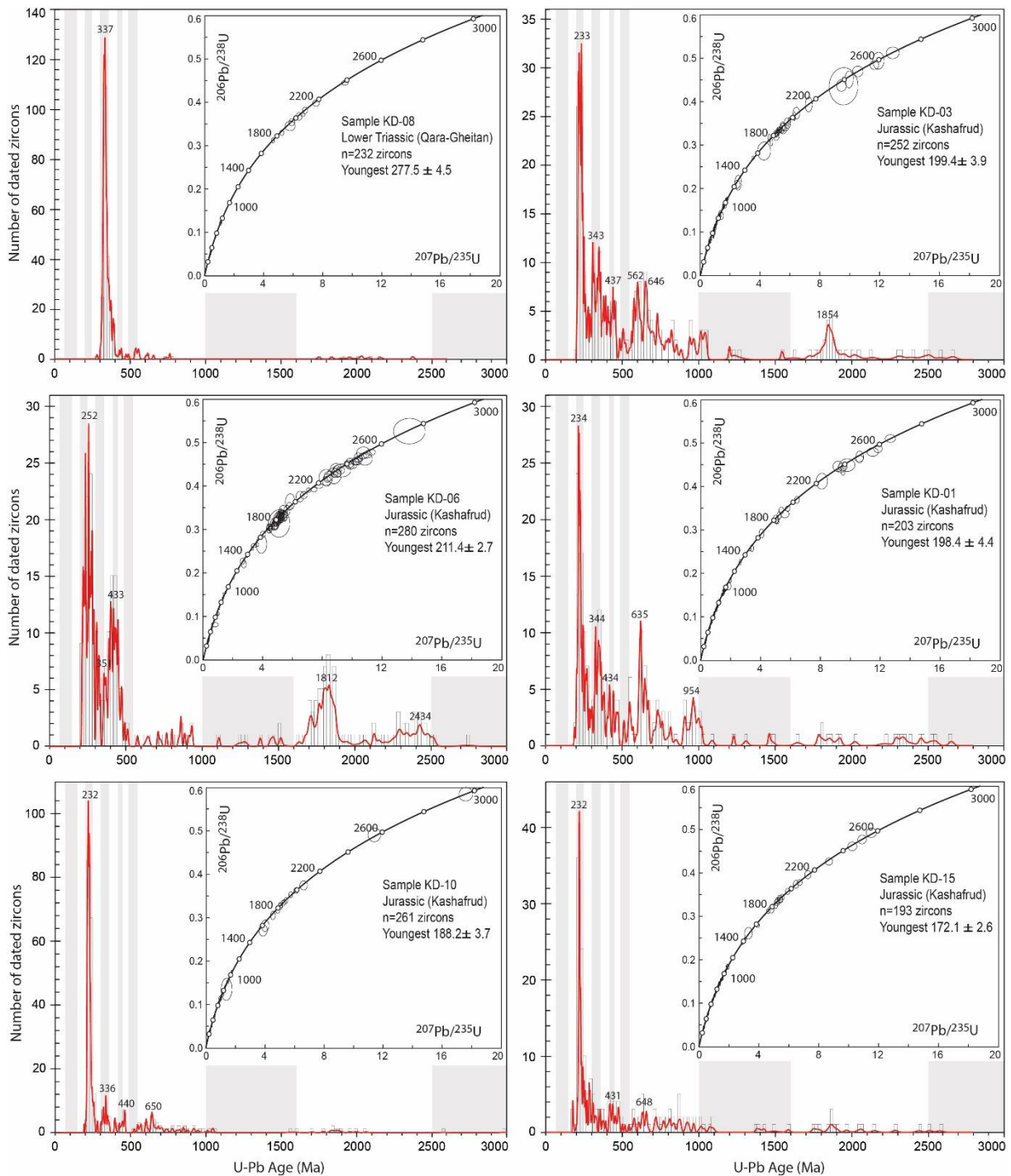


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1123 Fig. 6. Representative cathodoluminescence images of dated detrital zircons of the Kopet Dagh
 1124 sandstones. White circles show laser spot positions for U-Pb age analyses.

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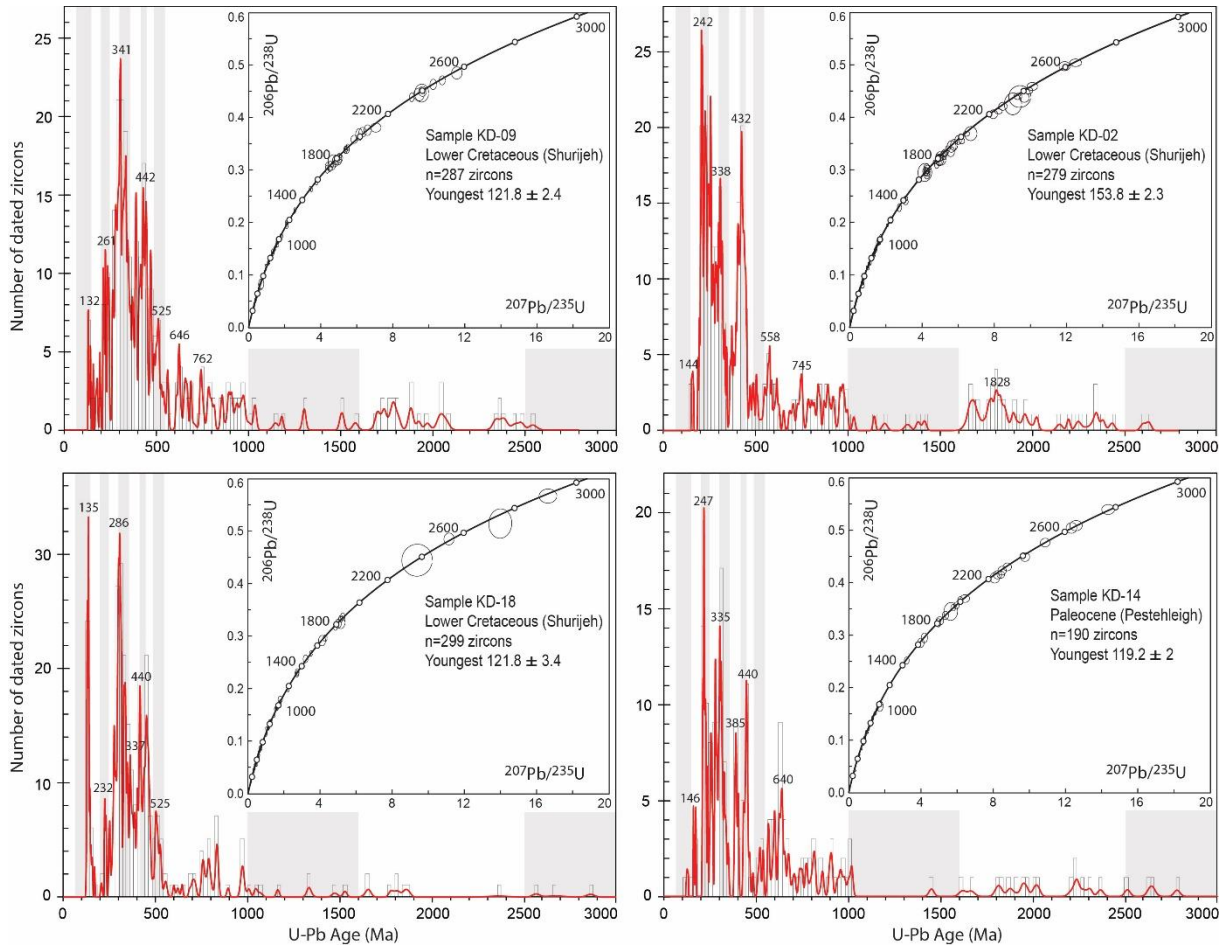


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1128 Fig. 7. Detrital zircon $^{206}\text{Pb}/^{238}\text{U}$ age probability density diagrams with corresponding
 1129 Concordia plots for the Lower Triassic and Jurassic samples KD-01, KD-03, KD-06, KD-08,
 1130 KD-10, and KD-15. Ages with discordance greater than 5% are excluded. Time scale after
 1131 [Gradstein and others \(2012\)](#).

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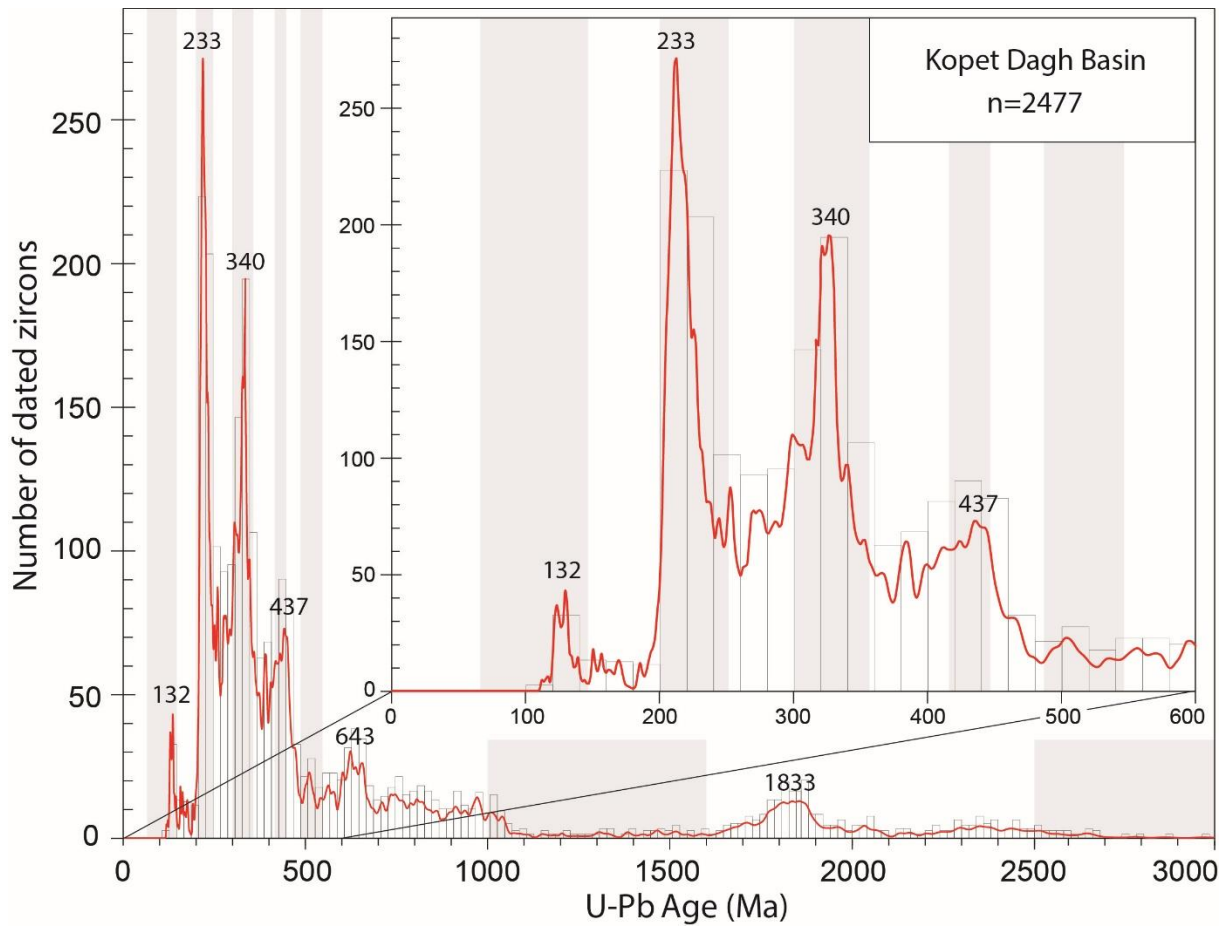


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1135 Fig. 8. Detrital zircon $^{206}\text{Pb}/^{238}\text{U}$ age population probability density diagrams with
1136 corresponding Concordia plots for Early Cretaceous and Paleocene samples KD-02, KD-09,
1137 KD-14 and KD-18. Ages with discordance greater than 5% are excluded. Time scale after
1138 [Gradstein and others \(2012\)](#).

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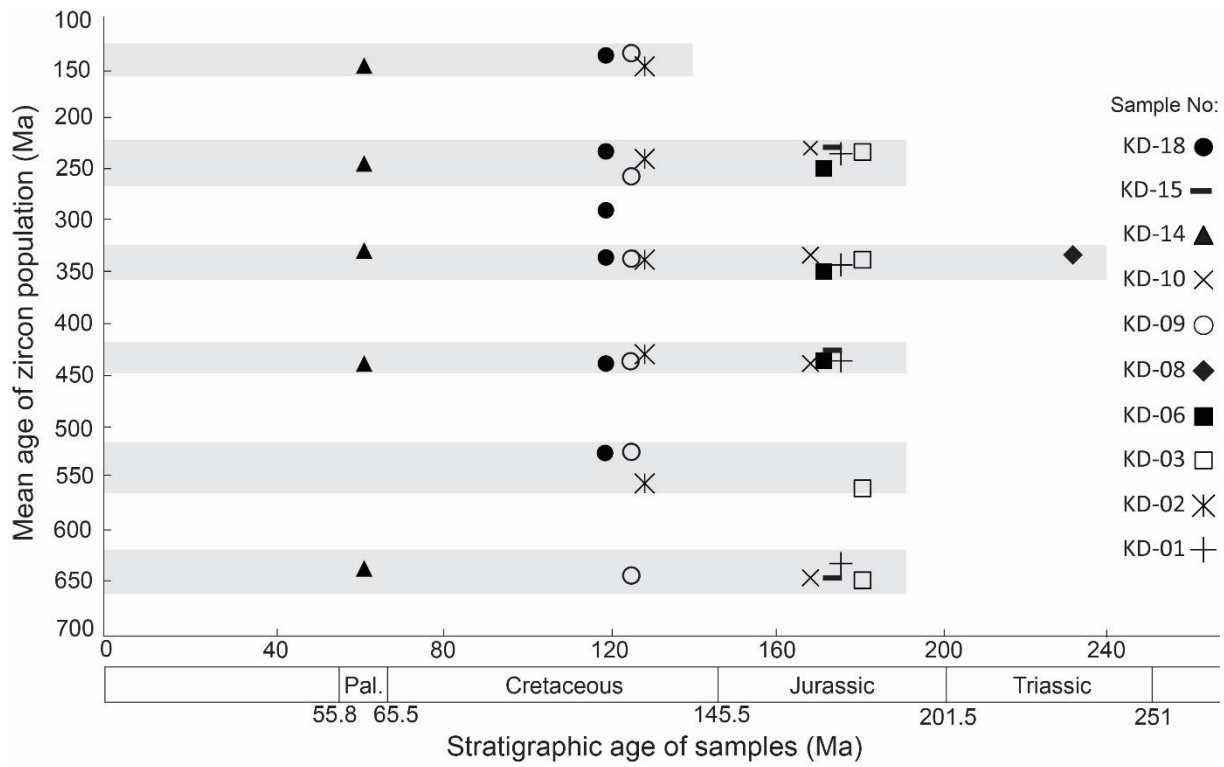
1142 Fig. 9. Combined zircon U-Pb age distribution pattern of Kopet Dagh sandstone samples.

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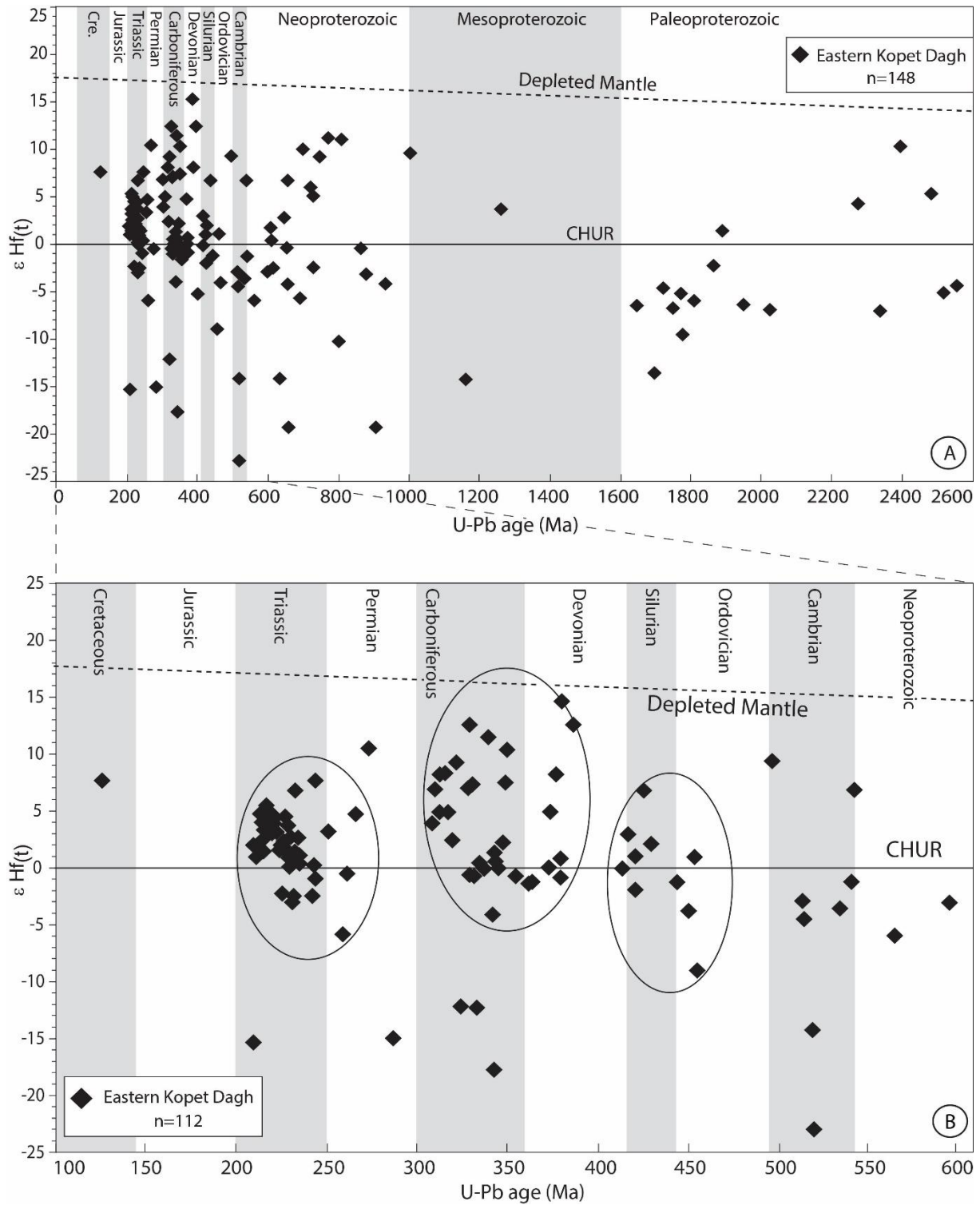
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1148 Fig. 10. Detrital zircon U-Pb mean age populations versus stratigraphic age of the Kopet Dagh

1149 sandstones. Time scale after [Gradstein and others \(2012\)](#).

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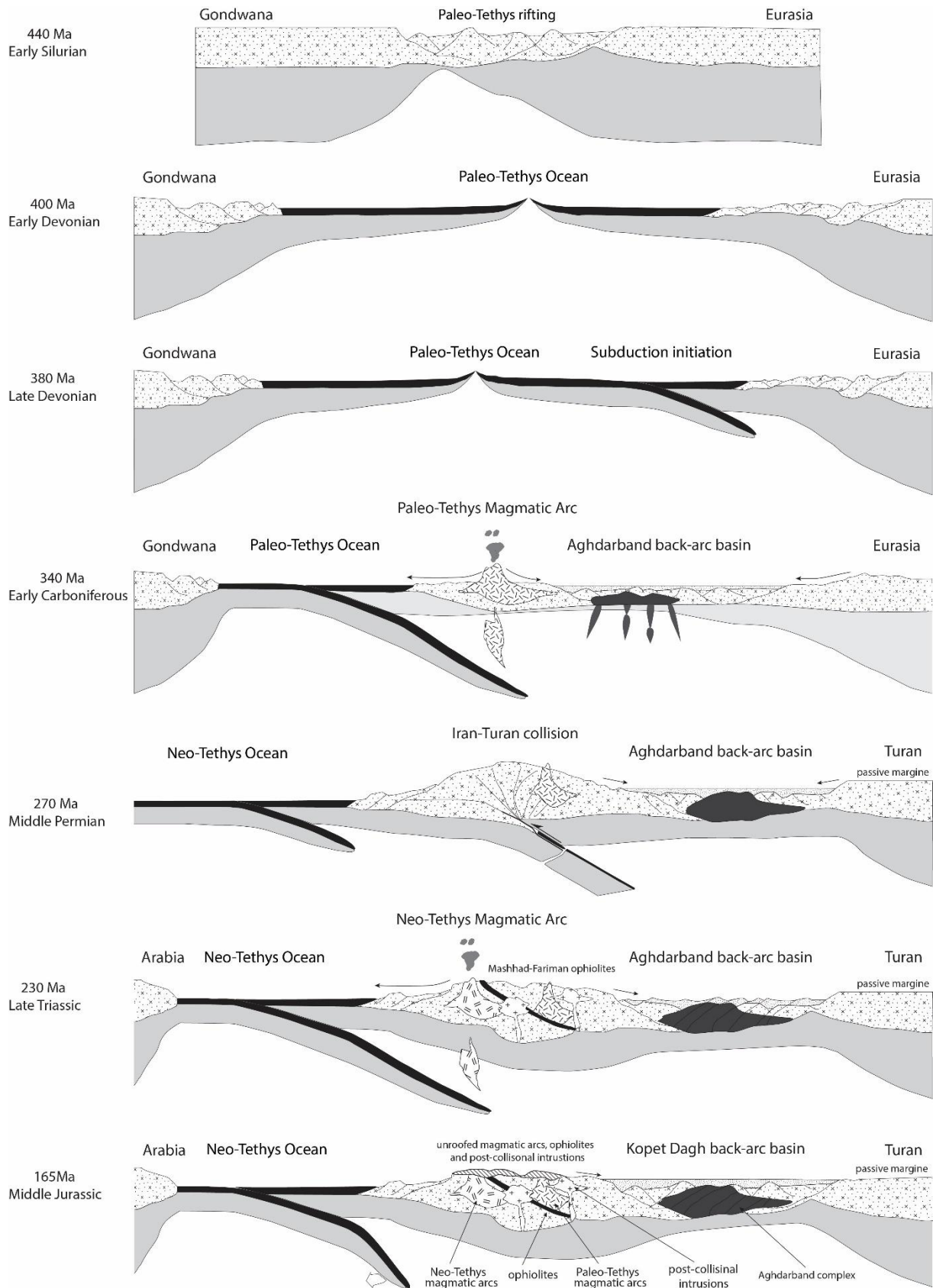
1153 Fig. 11. Time-corrected $\epsilon\text{-Hf}(t)$ values versus $^{206}\text{U}/^{238}\text{Pb}$ ages of Kopet Dagh detrital zircons.

1154 Depleted mantle evolution trend (dashed line) from Griffin and others (2000), and Chondritic

1155 Uniform Reservoir values (CHUR) from Bouvier and others (2008).

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1159 Fig. 12. Tectonic reconstruction along N-S profiles of the Kopet Dag Basin through time.

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