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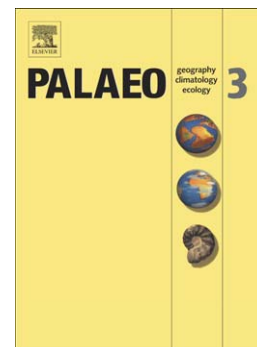
Arabia-Eurasia collision and the forcing of mid Cenozoic global cooling

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1 **Arabia-Eurasia collision and the forcing**
2 **of mid Cenozoic global cooling**

3

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7

8 **Abstract**

9 The end of the Eocene greenhouse world was the most dramatic phase in the
10 long-term cooling trend of the Cenozoic Era. Here we show that the Arabia-Eurasia
11 collision and the closure of the Tethys ocean gateway began in the Late Eocene at ~35
12 Ma, up to 25 million years earlier than in many reconstructions. We suggest that
13 global cooling was forced by processes associated with the initial collision that
14 reduced atmospheric CO₂. These are: 1) waning volcanism across southwest Asia; 2)
15 increased organic carbon storage in Paratethyan basins (e.g. Black Sea and South
16 Caspian); 3) increased silicate weathering in the collision zone and, 4) a shift towards
17 modern patterns of ocean currents, associated with increased vigour in circulation and
18 organic productivity.

19 *Keywords:* Eocene; Oligocene; Tethys; Arabia-Eurasia collision; global cooling.

20

21 **1. Introduction**

22 Stable isotopic data for the early Cenozoic (Paleocene to Eocene) show a long-
23 term pattern of cooling (Miller et al., 1987; Zachos et al., 2001; Tripathi et al., 2003)
24 followed by the rapid expansion of the Antarctic continental ice sheet in the latest
25 Eocene to earliest Oligocene (Ditchfield et al., 1994; Zachos et al., 2001). The latter

26 event, Oi-1, represents a 400 kyr-long glacial, initiated by reorganisation of the
27 ocean/climate system. This is evidenced by global shifts in the distribution of marine
28 biogenic sediments, including a ~1 km deepening of the calcite compensation depth
29 (CCD) (Coxall et al., 2005) and an overall increase in ocean fertility (Baldauf and
30 Barron, 1990; Salamy and Zachos, 1999; Thomas et al., 2000). A sharp positive
31 carbon isotope excursion (~0.5 ‰) indicates a significant perturbation in the global
32 carbon cycle (Zachos et al., 2001). High deep sea $\delta^{18}\text{O}$ values (~2.5 ‰) during this
33 event indicate permanent ice sheets, ~50% the size of the present day Antarctica ice
34 sheet (Zachos et al., 2001). Significant cool-water upwelling during Oi-1 (Kennett and
35 Barker, 1990; Barron et al., 1991; Diester-Haass, 1996; Salamy and Zachos, 1999;
36 Exon et al., 2002) is supported by a pattern of declining biotic diversity among marine
37 micro-invertebrates and dinoflagellates (Cifelli, 1969; Corliss, 1979; Benson et al.,
38 1984), diversification of the diatoms (Katz et al., 2004) and a widespread change from
39 carbonate (calcareous nannoplankton, foraminifers) to biosiliceous (diatom) oozes
40 along the Antarctic margin. Oi-1 also coincides with a shift in continental floral belts
41 (Frakes et al., 1992) and aridification and cooling in continental interiors Dupont-
42 Nivet et al., 2007; Zanazzi et al., 2007).

43 The causes of the Oi-1 glaciation remain contentious and have hitherto focused
44 on drivers from the southern high latitudes. Two first order causal hypotheses
45 dominate thinking on mid Cenozoic climate change: 1) opening of ocean gateways
46 separating Antarctica from other continents (Kennett, 1977); 2) reduction of
47 atmospheric CO_2 levels (DeConto and Pollard, 2003). Both hypotheses have caveats.
48 Recent models indicate that changes in oceanic heat transport as the result of
49 Antarctic isolation were too small to initiate Antarctic glaciation (Huber and Nof,
50 2006). Also, the precise timing of circum-Antarctic gateways is controversial (Pfuhl

51 and McCave, 2005; Scher and Martin, 2006; Livermore et al., 2007). End Eocene
52 decline in atmospheric CO₂ is supported by proxy data (Pagani et al., 2005), but this
53 leads to the question: what caused the decline?

54 Different lines of evidence indicates that initial collision of the Arabian and
55 Eurasian plates and closure of the Tethys Ocean took place at ~35 Ma (Late Eocene),
56 up to 25 million years earlier than in many plate tectonic or oceanographic
57 reconstructions (Woodruff and Savin, 1989; McQuarrie et al., 2003; Guest et al.,
58 2006), but consistent with geologic data from across the collision zone, used in other
59 reconstructions to argue for an Eocene age (Hempton, 1985; Vincent et al., 2005;
60 Jassim and Goff, 2006). This collision caused constriction of the Tethys Gateway,
61 which previously linked the Indian and Atlantic oceans (Fig. 1). We hypothesize that
62 this event caused large-scale, multiple feedbacks in the carbon cycle that promoted
63 global cooling and the Oi-1 glaciation.

64

65 **2. Date of initial Arabia-Eurasia continental collision**

66 There is considerable evidence for a Late Eocene (~35 Ma) age for the initial
67 Arabia-Eurasia collision and elimination of intervening Tethyan oceanic crust (Figs 2
68 and 3). Data include the timing of the following: compressional deformation, major
69 surface uplift, exhumation, non-deposition or angular unconformities; sediment
70 provenance switches and onset of terrestrial sedimentation, changes in
71 palaeobiogeography and the switch-off of arc magmatism. The data divide into
72 geographical sets on either side of the original plate suture; key regions are
73 summarised in Fig. 3. Note that Arabia was a promontory of the African plate before
74 the opening of the Red Sea in the mid Cenozoic, after initial Arabia-Eurasia collision.

75 To the south of the Arabia-Eurasia suture zone much of the collision history is
76 recorded in the tectono-stratigraphy of the Zagros Mountains in SW Iran and adjacent
77 parts of Iraq and Turkey. A regional Late Eocene – Early Oligocene angular
78 unconformity is recognised in the northeast of the Zagros (Hessami et al., 2001) (Fig.
79 3), interpreted by these authors as the early record of collision in an incipient foreland
80 basin. Over much of the Zagros, Oligocene deposition was dominated by shallow
81 marine carbonates of the Asmari Formation and its equivalents (Nadjafi et al., 2004),
82 but approaching the suture to the northeast the carbonates are replaced by sandstones
83 of the Razak Formation, shed from the region of the suture zone (Beydoun et al.,
84 1992). Close to the suture in southwest Iran, in the Kermanshah-Hamedan area, some
85 of the thrusts in the Zagros are post-Late Eocene to pre-Early Miocene, and are
86 unconformably overlain by Upper Oligocene – Lower Miocene conglomerates (Agard
87 et al., 2005) (Fig. 3). The thrust stack contains both Eocene volcanics and sedimentary
88 rocks of Eurasian affinity and Cretaceous sediments and ophiolites from the northeast
89 side of the Arabian plate (Agard et al., 2005).

90 In northeast Iraq, Upper Eocene terrestrial clastics of the Gercus Formation
91 unconformably overlie deformed Mesozoic strata (Dhannoun et al., 1988). These
92 strata and their underlying unconformity indicate compressional deformation and at
93 least local sub-aerial uplift and erosion of the northeast edge of the Arabian plate by
94 the Late Eocene, and have been interpreted as indicators of initial continental collision
95 at this time (Jassim and Goff, 2006).

96 At the eastern end of the collision zone in northern Oman, a record of stable
97 carbonate sedimentation from the latest Cretaceous – early Tertiary was terminated by
98 Late Oligocene – Miocene folding (Searle, 1988). Collectively, these data record
99 compressional deformation on the north Arabian margin from the Late Eocene

100 onwards (Fig. 3). Late Eocene compressional deformation also occurred at the
101 western side of the collision zone, from Syria at least as far west as Algeria (Guiraud
102 and Bosworth, 1999; Benaouali-Mebarek et al., 2006); it is not clear where effects of
103 the Arabia-Eurasia collision pass westwards in to the rather enigmatic “Atlas” phase
104 of deformation on the North African margin.

105 North of the suture, the Eurasian plate preserves a similar record of Late
106 Eocene – Oligocene compressional deformation, uplift and associated sedimentation
107 (Fig. 2). Close to the suture zone (Fig. 2), strata and igneous rocks as young as the
108 Middle Eocene were folded and thrust, in places onto the Arabian plate, before being
109 unconformably overlain by Oligocene sediments (Hempton, 1985; Yilmaz, 1993;
110 Yigitbas and Yilmaz, 1996; Agard et al., 2005). Late Eocene thrusting in the Kyrenia
111 Range of northern Cyprus is documented by deformed flysch and olistostrome
112 deposits of this age, overlain unconformably by Lower Oligocene conglomerates and
113 turbidites (Robertson and Woodcock, 1986). In the Berit region of southeast Turkey, a
114 mid Eocene to earliest Miocene melange incorporates material derived from the
115 Eurasian margin and is overlain by Lower Miocene turbidites, indicating that the
116 Arabian plate had underthrust Eurasia by the earliest Miocene (Robertson et al., 2004)
117 (Fig. 3). Within south-central Turkey several sedimentary basins, including Ulukışla
118 (Fig. 2), underwent Late Eocene compressional deformation, with folding, thrusting
119 and exhumation of volcanic rocks, turbidites and other sedimentary rocks deposited
120 during Paleocene – Middle Eocene extension (Clark and Robertson, 2005; Jaffey and
121 Robertson, 2005) (Fig. 3).

122 Eocene strata in the NW Greater Caucasus were deformed, exhumed and
123 eroded before the deposition of Oligocene clastics (Aleksin and Ratner, 1967)
124 indicating at least local deformation in this region near to the Eocene-Oligocene

125 boundary (Fig. 3). Parts of the western Greater Caucasus were emergent by at least
126 the Early Oligocene (Vincent et al., 2007). Upper Eocene olistostromes south of the
127 Greater Caucasus are interpreted as the result of compressional deformation (Banks et
128 al., 1997), while seismic data from the margins of the eastern Black Sea show
129 compressional deformation in the Late Eocene (Robinson et al., 1996). Syn-
130 sedimentary slumps accompanied deposition of Upper Eocene turbidites in the
131 Talysh, at the western margin of the South Caspian Basin (Vincent et al., 2005) (Fig.
132 3). These relatively fine-grained marine strata are overlain by a coarsening-upwards
133 Oligocene succession that includes boulder-scale conglomerates. This volcanic-free
134 stratigraphy superseded a pre-late Eocene deep marine succession with abundant
135 volcanism, including pillow basalts and tuffs. The Alborz range of northern Iran
136 switched from a Middle Eocene depocentre, including turbidites and tuffs, into an
137 emergent range by the early Oligocene (Stöcklin, 1974; Anell et al., 1975;
138 Alavi, 1996; Guest et al., 2006,) (Fig. 3). Late Eocene – Oligocene deformation
139 therefore occurred far to the north of the suture, suggesting that deformation
140 propagated rapidly into the interior of Eurasia at the time of initial plate collision
141 (Figs 2 and 3) (Robinson et al., 1996; Banks et al., 1997; Vincent et al., 2005; Vincent
142 et al., 2007).

143 A Late Eocene initial collision is consistent with faunal data. There was
144 progressive creation of separate Mediterranean and Indian Ocean marine realms, and
145 migration of Eurasian and African/Arabian non-marine faunas (Harzhauser et al.,
146 2002; Kappelman et al., 2003; Harzhauser et al., 2007). This is demonstrated by the
147 tridacnine and strombid bivalves (Harzhauser et al., 2007), which show
148 biogeographical divergence in the Oligocene. Gastropod assemblages also define two
149 separate Tethys sub-provinces during the Oligocene, with an ill-defined boundary

150 within Iran and a rapid increase in endemism in the early Miocene (Harzhauser et al.,
151 2002). The influx of Eurasian mammals into Africa indicates a land connection
152 between Africa-Arabia and Eurasia existed by the Oligocene-Miocene boundary
153 (Kappelman et al., 2003).

154 Tethyan sections at the Eocene-Oligocene transition show coeval faunal
155 overturn in benthic foraminifera, accompanied by decreasing ventilation, preceding an
156 increased intensity of abyssal circulation associated with the initial entry of bottom
157 waters (likely to be North Atlantic Deep Water, NADW) and bolivinid/uvigerinid
158 planktonic foraminifera blooms along the northern Tethys margin (Barbieri et al.,
159 2003).

160

161 **3. Collision, the carbon cycle and oceanography**

162 Late Eocene closure of Tethys was coincident with declining $p\text{CO}_2$ levels
163 (Pagani et al., 2005), implicated as a major driver for global cooling and Antarctic
164 glaciation (DeConto and Pollard, 2003). We propose four potential mechanisms for
165 reducing $p\text{CO}_2$ associated with initial Arabia-Eurasia collision and its effects on
166 carbon fluxes and/or oceanographic circulation: decline of arc magmatism; storage of
167 organic carbon in sedimentary basins; increased silicate weathering; stimulation of
168 more vigorous, meridional ocean currents.

169

170 *3.1. Declining Eocene arc magmatism in southwest Eurasia.*

171 Before the Arabia-Eurasia collision the Eurasian continental margin
172 experienced arc magmatism as the result of the northwards subduction of Tethyan
173 (strictly, Neo-Tethyan) oceanic crust. This magmatism provides a time constraint on
174 the maximum likely age for initial continental collision, and would have been a net

175 source of atmospheric CO₂. Across much of Iran and Turkey and adjacent areas there
176 was a highly productive magmatic arc/back-arc system between ~50 and ~35 Ma.
177 Magmatism was coincident with the renewal of northern motion of Africa-Arabia
178 with respect to Eurasia, after a hiatus between 75 and 49 Ma (Dewey et al., 1989).
179 Peak magmatism occurred in the Middle Eocene, close to 40 Ma, at which time
180 volcanic successions accumulated at a rate of ~1.8 mm/yr, reached 4-8 km in
181 thickness and occurred across an area of >2 million km² (Amidi et al., 1984; Kazmin
182 et al., 1986; Brunet et al., 2003; McQuarrie et al., 2003; Ramezani and Tucker, 2003;
183 Alpaslan et al., 2004; Vincent et al., 2005; Arslan and Aslan, 2006; Fig. 4). In detail,
184 at least 4 km of intermediate-acidic volcanics are intercalated with mid-Eocene
185 Nummulitic limestones in the Urumieh-Dokhtar arc in Iran (Berberian et al., 1982).
186 Eight kilometres of mainly Middle Eocene volcanics and volcanogenic turbidites are
187 recorded from the Talysh, adjacent to the South Caspian Basin (Vincent et al., 2005).
188 Five km of Eocene andesitic volcanics and deep water clastics were deposited in the
189 Alborz Mountains (Stöcklin, 1974; Alavi, 1996). Volcanism waned in the Late
190 Eocene and there was little activity in the Oligocene (Fig. 4), though minor and
191 sporadic magmatism has continued to the present day over much of the collision zone
192 (Pearce et al., 1990).

193 Declining arc magmatism in the Late Eocene is consistent with the early
194 deformation history of the collision zone (Fig. 2), whereby Late Eocene initial
195 collision of the Arabian and Eurasian plates terminated oceanic subduction, ended
196 back-arc continental extension across southwest Asia (Vincent et al., 2005) and
197 generated compressional deformation and surface uplift. Abundant Middle Eocene arc
198 magmatism across SW Asia would have promoted high atmospheric CO₂ levels,
199 although the precise amount is not known. This highly productive arc coincides with

200 the Middle Eocene climatic optimum, previously attributed to an unspecified rise in
201 ridge or arc magmatism (Bohaty and Zachos, 2003). Conversely, the sharp reduction
202 in arc magmatism, brought about by initial Arabia-Eurasia collision, would have
203 reduced CO₂ degassing into the atmosphere, and so acted to reduce global
204 temperatures.

205

206 *3.2. Isolation of Paratethys and organic carbon storage.*

207 A new oceanographic configuration formed between the Alps and the Aral Sea
208 during the Late Eocene and Oligocene (Veto, 1987; Jones and Simmons, 1997; Rögl,
209 1999; Fig. 4). The basins were isolated from the global circulation, were prone to
210 anoxia, and are collectively referred to as Paratethys or the Paratethyan basins. In the
211 South Caspian and Black Sea basins the depocentres were located over blocks of
212 highly attenuated continental crust or even oceanic crust (Finetti et al., 1988; Mangino
213 and Priestley, 1998). These basement blocks are products of Mesozoic or early
214 Cenozoic extension across southwestern Asia. Upper Eocene and Oligocene strata are
215 commonly mud-prone and organic-rich across the region (Robinson et al., 1996;
216 Vincent et al., 2005). Such organic-rich mudrocks are the main hydrocarbon source
217 rock for the prolific oil fields of the Carpathians and South Caspian Basin, and are the
218 main potential source rock in the eastern Black Sea. Total organic carbon (TOC)
219 values reach 14% for the 2000 m thick Maykop Suite in the South Caspian Basin
220 (Robinson et al., 1996; Katz et al., 2000). In the ~1000 m thick coeval strata of the
221 Greater Caucasus, estimated average TOC values are ~1.5 to 2%. Typical thicknesses
222 for the age equivalent Menilite Formation in eastern Europe are ~300 m, with average
223 TOC of 2% (Veto, 1987). Based on these estimates of stratal thicknesses, extents and

224 average TOC, we estimate total organic sedimentary carbon in the combined Maykop
225 and Menilite units at 60×10^{12} T.

226 Our estimate for organic carbon stored in the uppermost Eocene-Oligocene
227 strata of the Paratethyan basins corresponds to an average deposition rate of $\sim 6 \times 10^{12}$
228 T per Ma through this interval, equivalent to $\sim 12\%$ of the estimated global organic
229 carbon flux in the late Paleogene (Raymo, 1994). This flux is a crude estimate, given
230 that the distribution of organic carbon within the succession is poorly known but
231 unlikely to be even. The overall effect of the carbon drawdown would have
232 suppressed atmospheric CO_2 levels throughout the latest Eocene and Oligocene.

233

234 *3.3. Increased silicate weathering.*

235 Continental collision and increased sub-aerial erosion in newly elevated areas
236 would enhance low latitude silicate weathering (Raymo and Ruddiman, 1992), which
237 in turn promotes CO_2 drawdown from the atmosphere by reactions that can be
238 summarised as:

239



241

242 Evidence for exposure and increased erosion comes from the presence of non-
243 marine clastics or uplifted areas across large parts of the Arabia-Eurasia collision zone
244 from the Late Eocene onwards. The precise contribution to global CO_2 drawdown
245 from silicate weathering in the collision zone is difficult to quantify, and likely to
246 have been small given the area and likely rates involved when compared with global
247 rates, but it acted in the right sense to promote climatic cooling. Enhanced weathering

248 and erosion could also help account for the increase in the oceanic $^{87}\text{Sr}/^{86}\text{Sr}$ in the
249 Late Eocene (Richter et al., 1992; Mead and Hodell, 1995).

250

251 *3.4. Oceanographic changes.*

252 Closure of Tethys resulted in a restructuring of Indian and Atlantic Ocean
253 currents, closer to a modern pattern of ocean circulation and upwelling (Fig. 1). In the
254 Cretaceous to Eocene (the “Proteus Ocean” of Kennett and Barker, 1990) low latitude
255 surface currents were dominated by the circum-global westwards flow from the Indian
256 Ocean to the Pacific via the Tethys and Panama gateways (Bush, 1997; Hallam, 1969;
257 Huber and Sloan, 2001; Fig. 1). At about 37.5 Ma circum-equatorial surface water
258 was directed southwards in the Indian Ocean via the Agulhas Current, as a result of
259 the constricting Tethys Gateway (Diekmann et al., 2004). This current is a possible
260 source of the moisture thought to be a critical element in maintaining a large mid
261 Cenozoic Antarctic ice sheet (Zachos et al., 2001). Within the western Tethys
262 (Mediterranean) region there was an increased intensity of abyssal circulation
263 associated with the initial entry of NADW across the Eocene-Oligocene transition
264 (Barbieri et al., 2003). Influx of cold corrosive deep water at ~34 Ma was a likely
265 cause of marked faunal overturn in benthic foraminifera (Coccioni and Galeotti,
266 2003). Contourite deposition began in Cyprus at ~36 Ma (Kahler and Stow, 1998),
267 also indicating increased ocean current vigour.

268 Stable and Nd isotope data show that a marine connection between the Indian
269 and Atlantic oceans persisted into the Miocene (Woodruff and Savin, 1989; Stille et
270 al., 1996), but as argued here, this seaway cannot have been floored by oceanic crust.

271 Tethys closure was just one aspect of mid Cenozoic plate re-configuration and
272 oceanographic change. The widening North Atlantic led to the start of NADW at ~35

273 Ma (Wold, 1994; Zachos et al., 2001; Via and Thomas, 2006). Atlantic circulation
274 patterns similar to the present day were established at this time (Via and Thomas,
275 2006). Although the precise timing for the opening of Antarctic gateways is still
276 debated, the trend towards isolation is clear in plate reconstructions (Livermore et al.,
277 2007). Likewise, Mediterranean tectonics involved rapid compressional and
278 extensional events in the early Cenozoic, in the context of the overall convergence of
279 Africa and Europe (Dewey et al., 1989; Rubatto et al., 1998), but without complete
280 severance of the Tethyan seaway west of Arabia.

281 Oceanographic changes have been implicated in global climate change via
282 increased upwelling, organic productivity and hence atmospheric CO₂ drawdown
283 (Diester-Haass and Zahn, 1996, 2001; Schumacher and Lazarus, 2004; Anderson and
284 Delaney, 2005). Our point is that Late Eocene Tethys closure is a previously
285 unappreciated factor in this global re-organisation.

286

287 **4. Conclusions**

288 Oceanographic, plate tectonic and climatic modelling studies commonly take
289 ~14 to 10 Ma (mid Miocene) as both the end of the Tethys connection between the
290 Indian and Atlantic oceans and the initial Arabia-Eurasia collision (Woodruff and
291 Savin, 1989; McQuarrie et al., 2003). Our interpretation of the collision is that the last
292 oceanic plate separation between Arabia and Eurasia was in the Late Eocene at ~35
293 Ma (Fig. 1), agreeing with previous estimates for this age based on geological patterns
294 within the collision zone (Jassim and Goff, 2006; Vincent et al., 2007).

295 Initial Arabia-Eurasia plate collision and closure of the Tethys Ocean provides
296 four complementary mechanisms for reducing atmospheric CO₂ and global cooling:
297 1) the waning of pre-collision arc magmatism, 2) storage of organic carbon in the

298 Paratethyan basins, 3) an increase in silicate weathering, 4) re-organisation of ocean
299 currents, consistent with global end Eocene increases in ocean current vigour, organic
300 productivity and hence CO₂ drawdown. We contend that all these mechanisms acted
301 together to help take the Earth across a threshold into the icehouse world at the Oi-1
302 event.

303

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308

309 **Figures**

310 Fig. 1. Palaeogeographic and oceanographic reconstructions before and after the
311 demise of the Tethys Ocean gateway. **(A)** Eocene period, with westerly transport of
312 warm Indian Ocean water into the Atlantic via Tethys. **(B)** Oligocene, with
313 connection between the Indian and Atlantic oceans impeded by the Arabia-Eurasia
314 collision zone. Ocean currents derived from Bush (1997); Diekmann et al. (2004);
315 Kennett and Barker (1990); Stille et al. (1996); Thomas et al. (2003); Via and Thomas
316 (2006); von der Heydt and Dijkstra (2006).

317

318 Fig. 2. Present topography of the Arabia-Eurasia collision, location map for regions
319 summarised in Fig. 3, and position of the Arabia-Eurasia suture.

320

321 Fig. 3. Summary tectonostratigraphy for localities showing Late Eocene – Oligocene
322 deformation and/or uplift. Localities shown on Fig. 2. Derived from: Stöcklin, (1974);

323 Annells et al., (1975); Searle, (1988); Banks et al., (1997); Beydoun et al., (1992);
324 Hessami et al., (2001); Agard et al., (2005); Clark and Robertson, (2005); Vincent et
325 al., (2005, 2007); Guest et al., (2006); Boulton and Robertson, (2006); Jassim and
326 Goff, (2006); Robertson et al., (2006).

327

328 Fig. 4. Comparison of the present distribution of (A) Eocene and (B) Oligocene
329 magmatic rocks across southwest Asia. Derived from principally from Emami et al.,
330 (1993); Şenel (2002). Other sources summarised in Vincent et al. (2005). (B) also
331 shows the extent of Oligocene sediments from the Paratethyan basins (Veto, 1987).

332

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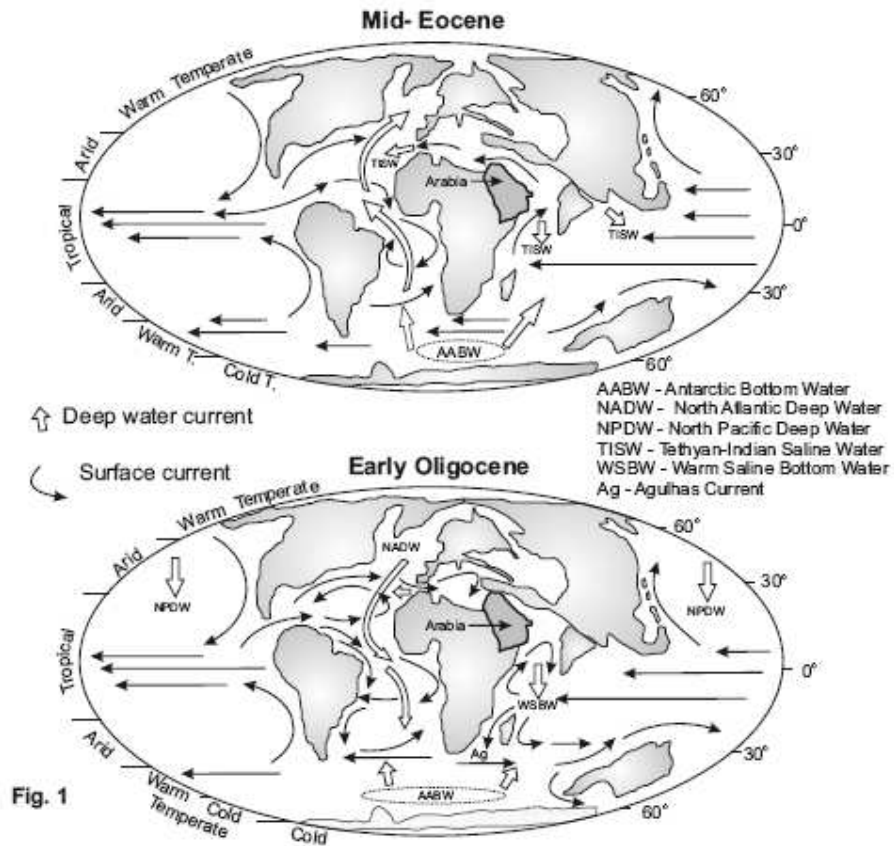
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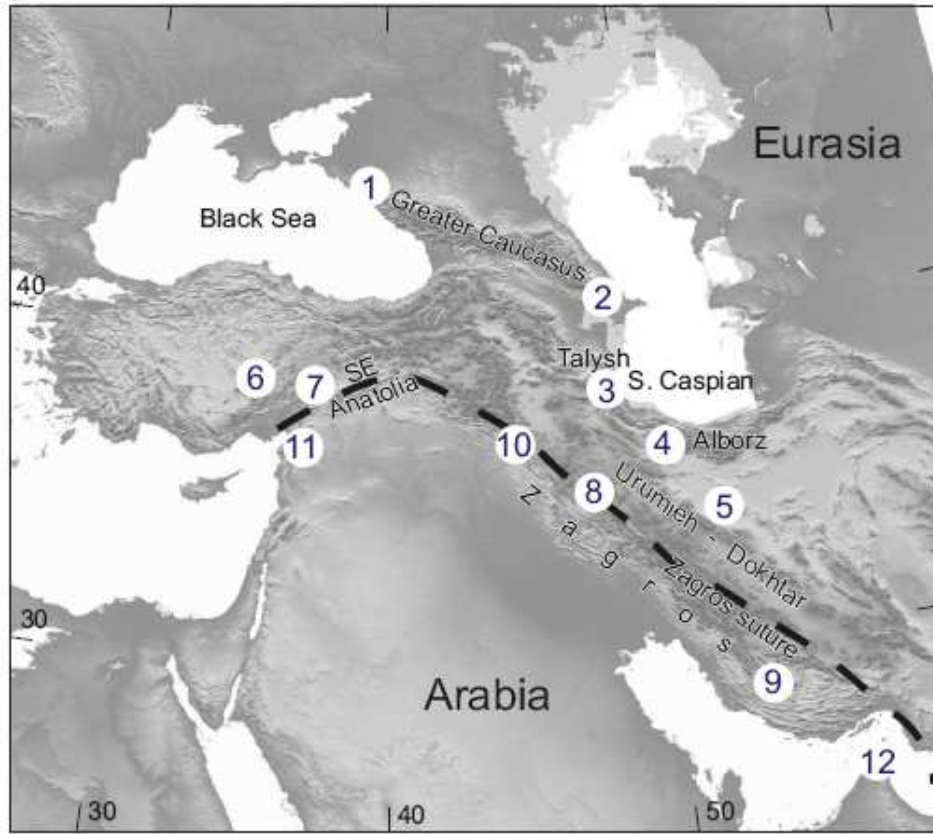


Fig. 2

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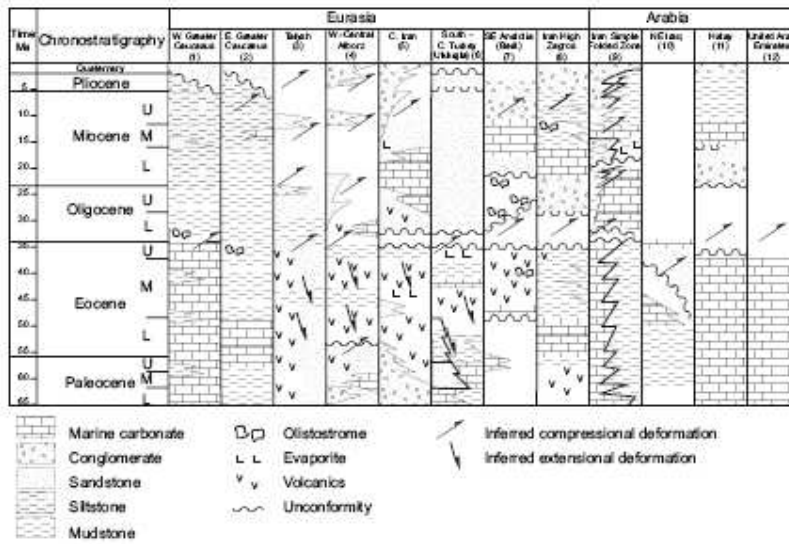
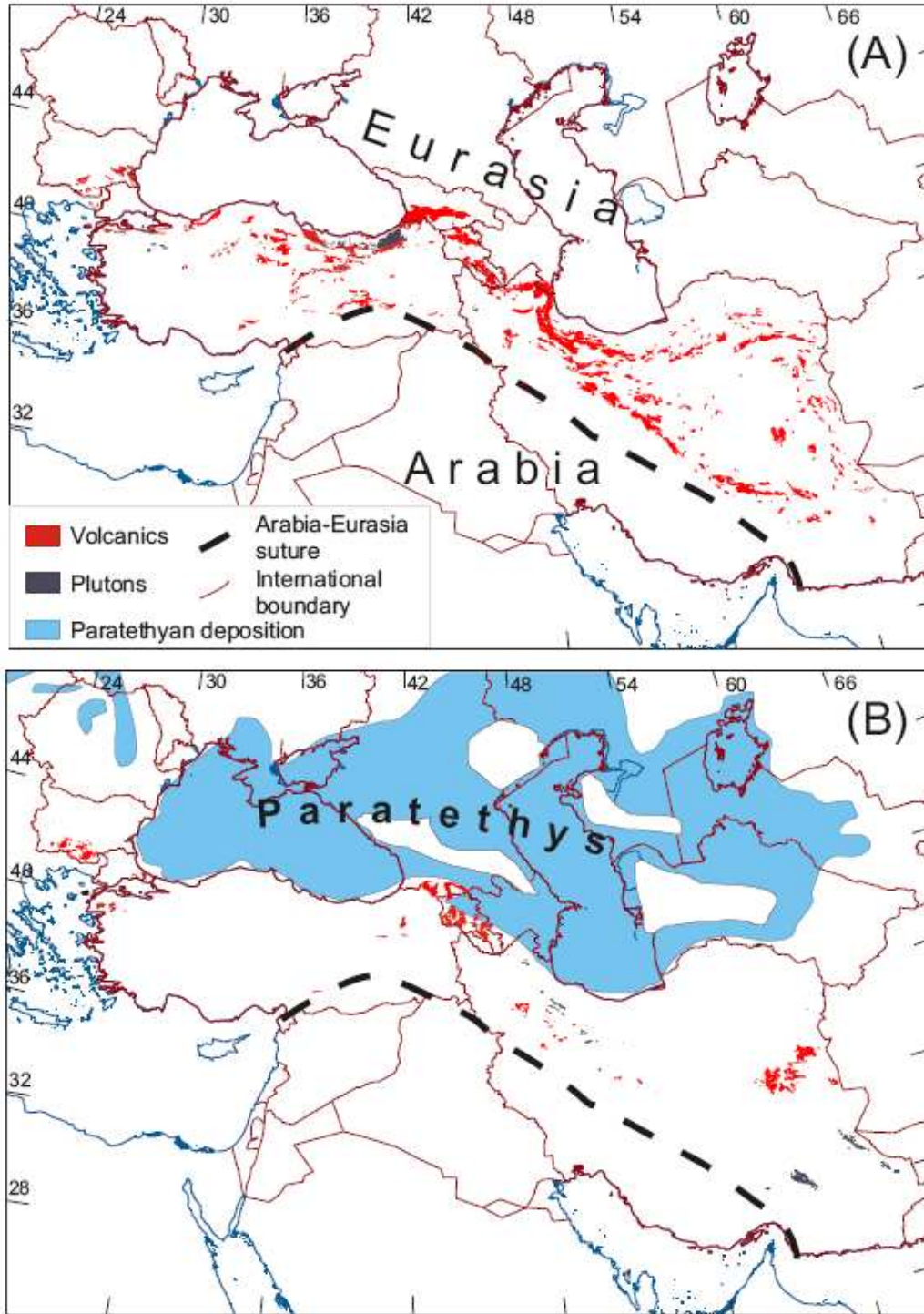


Fig. 3



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Fig. 4