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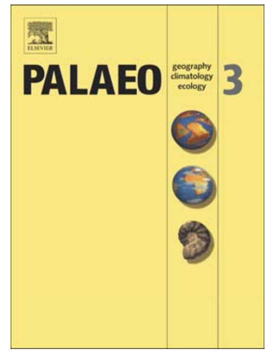

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Arabia-Eurasia collision and the forcing of mid Cenozoic global cooling

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Abstract

The end of the Eocene greenhouse world was the most dramatic phase in the long-term cooling trend of the Cenozoic Era. Here we show that the Arabia-Eurasia collision and the closure of the Tethys ocean gateway began in the Late Eocene at ~35 Ma, up to 25 million years earlier than in many reconstructions. We suggest that global cooling was forced by processes associated with the initial collision that reduced atmospheric CO_2. These are: 1) waning volcanism across southwest Asia; 2) increased organic carbon storage in Paratethyan basins (e.g. Black Sea and South Caspian); 3) increased silicate weathering in the collision zone and, 4) a shift towards modern patterns of ocean currents, associated with increased vigour in circulation and organic productivity.

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

1. Introduction

Stable isotopic data for the early Cenozoic (Paleocene to Eocene) show a long-term pattern of cooling (Miller et al., 1987; Zachos et al., 2001; Tripati et al., 2003) followed by the rapid expansion of the Antarctic continental ice sheet in the latest Eocene to earliest Oligocene (Ditchfield et al., 1994; Zachos et al., 2001). The latter event, Oi-1, represents a 400 kyr-long glacial, initiated by reorganisation of the ocean/climate system. This is evidenced by global shifts in the distribution of marine biogenic sediments, including a ~1 km deepening of the calcite compensation depth (CCD) (Coxall et al., 2005) and an overall increase in ocean fertility (Baldauf and Barron, 1990; Salamy and Zachos, 1999; Thomas et al., 2000). A sharp positive carbon isotope excursion (~0.5 ‰) indicates a significant perturbation in the global carbon cycle (Zachos et al., 2001). High deep sea $\delta^{18}O$ values (~2.5 ‰) during this event indicate permanent ice sheets, ~50% the size of the present day Antarctica ice sheet (Zachos et al., 2001). Significant cool-water upwelling during Oi-1 (Kennett and Barker, 1990; Barron et al., 1991; Diester-Haass, 1996; Salamy and Zachos, 1999; Exon et al., 2002) is supported by a pattern of declining biotic diversity among marine micro-invertebrates and dinoflagellates (Cifelli, 1969; Corliss, 1979; Benson et al., 1984), diversification of the diatoms (Katz et al., 2004) and a widespread change from carbonate (calcareous nannoplankton, foraminifers) to biosiliceous (diatom) oozes along the Antarctic margin. Oi-1 also coincides with a shift in continental floral belts (Frakes et al., 1992) and aridification and cooling in continental interiors Dupont-Nivet et al., 2007; Zanazzi et al., 2007).

The causes of the Oi-1 glaciation remain contentious and have hitherto focused on drivers from the southern high latitudes. Two first order causal hypotheses dominate thinking on mid Cenozoic climate change: 1) opening of ocean gateways separating Antarctica from other continents (Kennett, 1977); 2) reduction of atmospheric CO_2 levels (DeConto and Pollard, 2003). Both hypotheses have caveats. Recent models indicate that changes in oceanic heat transport as the result of Antarctic isolation were too small to initiate Antarctic glaciation (Huber and Nof, 2006). Also, the precise timing of circum-Antarctic gateways is controversial (Pfuhl and McCave, 2005; Scher and Martin, 2006: Livermore et al., 2007). End Eocene decline in atmospheric CO_2 is supported by proxy data (Pagani et al., 2005), but this leads to the question: what caused the decline?

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

2. Date of initial Arabia-Eurasia continental collision

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

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

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

At the eastern end of the collision zone in northern Oman, a record of stable carbonate sedimentation from the latest Cretaceous – early Tertiary was terminated by Late Oligocene – Miocene folding (Searle, 1988). Collectively, these data record compressional deformation on the north Arabian margin from the Late Eocene onwards (Fig. 3). Late Eocene compressional deformation also occurred at the western side of the collision zone, from Syria at least as far west as Algeria (Guiraud and Bosworth, 1999; Benaouali-Mebarek et al., 2006); it is not clear where effects of the Arabia-Eurasia collision pass westwards in to the rather enigmatic "Atlas" phase of deformation on the North African margin.

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

Eocene strata in the NW Greater Caucasus were deformed, exhumed and eroded before the deposition of Oligocene clastics (Aleksin and Ratner, 1967) indicating at least local deformation in this region near to the Eocene-Oligocene boundary (Fig. 3). Parts of the western Greater Caucasus were emergent by at least the Early Oligocene (Vincent et al., 2007). Upper Eocene olistostromes south of the Greater Caucasus are interpreted as the result of compressional deformation (Banks et al., 1997), while seismic data from the margins of the eastern Black Sea show compressional deformation in the Late Eocene (Robinson et al., 1996). Syn-sedimentary slumps accompanied deposition of Upper Eocene turbidites in the Talysh, at the western margin of the South Caspian Basin (Vincent et al., 2005) (Fig. 3). These relatively fine-grained marine strata are overlain by a coarsening-upwards Oligocene succession that includes boulder-scale conglomerates. This volcanic-free stratigraphy superseded a pre-late Eocene deep marine succession with abundant volcanism, including pillow basalts and tuffs. The Alborz range of northern Iran switched from a Middle Eocene depocentre, including turbidites and tuffs, into an emergent range by the early Oligocene (Stöcklin, 1974; Annell et al., 1975; Alavi,1996; Guest et al., 2006,) (Fig. 3). Late Eocene – Oligocene deformation therefore occurred far to the north of the suture, suggesting that deformation propagated rapidly into the interior of Eurasia at the time of initial plate collision (Figs 2 and 3) (Robinson et al., 1996; Banks et al., 1997; Vincent et al., 2005; Vincent et al., 2007).

A Late Eocene initial collision is consistent with faunal data. There was progressive creation of separate Mediterranean and Indian Ocean marine realms, and migration of Eurasian and African/Arabian non-marine faunas (Harzhauser et al., 2002; Kappelman et al., 2003; Harzhauser et al., 2007). This is demonstrated by the tridacnine and strombid bivalves (Harzhauser et al., 2007), which show biogeographical divergence in the Oligocene. Gastropod assemblages also define two separate Tethys sub-provinces during the Oligocene, with an ill-defined boundary within Iran and a rapid increase in endemism in the early Miocene (Harzhauser et al., 2002). The influx of Eurasian mammals into Africa indicates a land connection between Africa-Arabia and Eurasia existed by the Oligocene-Miocene boundary (Kappelman et al., 2003).

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

3. Collision, the carbon cycle and oceanography

Late Eocene closure of Tethys was coincident with declining pCO_2 levels (Pagani et al., 2005), implicated as a major driver for global cooling and Antarctic glaciation (DeConto and Pollard, 2003). We propose four potential mechanisms for reducing pCO_2 associated with initial Arabia-Eurasia collision and its effects on carbon fluxes and/or oceanographic circulation: decline of arc magmatism; storage of organic carbon in sedimentary basins; increased silicate weathering; stimulation of more vigorous, meridional ocean currents.

3.1. Declining Eocene arc magmatism in southwest Eurasia.

Before the Arabia-Eurasia collision the Eurasian continental margin experienced arc magmatism as the result of the northwards subduction of Tethyan (strictly, Neo-Tethyan) oceanic crust. This magmatism provides a time constraint on the maximum likely age for initial continental collision, and would have been a net source of atmospheric CO_2. Across much of Iran and Turkey and adjacent areas there was a highly productive magmatic arc/back-arc system between ~50 and ~35 Ma. Magmatism was coincident with the renewal of northern motion of Africa-Arabia with respect to Eurasia, after a hiatus between 75 and 49 Ma (Dewey et al., 1989). Peak magmatism occurred in the Middle Eocene, close to 40 Ma, at which time volcanic successions accumulated at a rate of ~1.8 mm/yr, reached 4-8 km in thickness and occurred across an area of >2 million km^2 (Amidi et al., 1984; Kazmin et al., 1986; Brunet et al., 2003; McQuarrie et al., 2003; Ramezani and Tucker, 2003; Alpaslan et al., 2004; Vincent et al., 2005; Arslan and Aslan, 2006; Fig. 4). In detail, at least 4 km of intermediate-acidic volcanics are intercalated with mid-Eocene Nummulitic limestones in the Urumieh-Dokhtar arc in Iran (Berberian et al., 1982). Eight kilometres of mainly Middle Eocene volcanics and volcanigenic turbidites are recorded from the Talysh, adjacent to the South Caspian Basin (Vincent et al., 2005). Five km of Eocene andesitic volcanics and deep water clastics were deposited in the Alborz Mountains (Stöcklin, 1974; Alavi, 1996). Volcanism waned in the Late Eocene and there was little activity in the Oligocene (Fig. 4), though minor and sporadic magmatism has continued to the present day over much of the collision zone (Pearce et al., 1990).

Declining arc magmatism in the Late Eocene is consisitent with the early deformation history of the collision zone (Fig. 2), whereby Late Eocene initial collision of the Arabian and Eurasian plates terminated oceanic subduction, ended back-arc continental extension across southwest Asia (Vincent et al., 2005) and generated compressional deformation and surface uplift. Abundant Middle Eocene arc magmatism across SW Asia would have promoted high atmospheric CO_2 levels, although the precise amount is not known. This highly productive arc coincides with the Middle Eocene climatic optimum, previously attributed to an unspecified rise in ridge or arc magmatism (Bohaty and Zachos, 2003). Conversely, the sharp reduction in arc magmatism, brought about by initial Arabia-Eurasia collision, would have reduced CO_2 degassing into the atmosphere, and so acted to reduce global temperatures.

3.2. Isolation of Paratethys and organic carbon storage.

A new oceanographic configuration formed between the Alps and the Aral Sea during the Late Eocene and Oligocene (Veto, 1987; Jones and Simmons, 1997; Rögl, 1999; Fig. 4). The basins were isolated from the global circulation, were prone to anoxia, and are collectively referred to as Paratethys or the Paratethyan basins. In the South Caspian and Black Sea basins the depocentres were located over blocks of highly attenuated continental crust or even oceanic crust (Finetti et al., 1988; Mangino and Priestley, 1998). These basement blocks are products of Mesozoic or early Cenozoic extension across southwestern Asia. Upper Eocene and Oligocene strata are commonly mud-prone and organic-rich across the region (Robinson et al., 1996; Vincent et al., 2005). Such organic-rich mudrocks are the main hydrocarbon source rock for the prolific oil fields of the Carpathians and South Caspian Basin, and are the main potential source rock in the eastern Black Sea. Total organic carbon (TOC) values reach 14% for the 2000 m thick Maykop Suite in the South Caspian Basin (Robinson et al., 1996; Katz et al., 2000). In the ~1000 m thick coeval strata of the Greater Caucasus, estimated average TOC values are ~1.5 to 2%. Typical thicknesses for the age equivalent Menilite Formation in eastern Europe are ~300 m, with average TOC of 2% (Veto, 1987). Based on these estimates of stratal thicknesses, extents and average TOC, we estimate total organic sedimentary carbon in the combined Maykop and Menilite units at 60 x 10^{12} T.

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

3.3. Increased silicate weathering.

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

$$CO_2 + CaSiO_3 \rightarrow CaCO_3 + SiO_2$$

Evidence for exposure and increased erosion comes from the presence of non-marine clastics or uplifted areas across large parts of the Arabia-Eurasia collision zone from the Late Eocene onwards. The precise contribution to global CO_2 drawdown from silicate weathering in the collision zone is difficult to quantify, and likely to have been small given the area and likely rates involved when compared with global rates, but it acted in the right sense to promote climatic cooling. Enhanced weathering and erosion could also help account for the increase in the oceanic $^{87}Sr/^{86}Sr$ in the Late Eocene (Richter et al., 1992; Mead and Hodell, 1995).

3.4. Oceanographic changes.

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

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

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

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

Oceanographic changes have been implicated in global climate change via increased upwelling, organic productivity and hence atmospheric CO_2 drawdown (Diester-Haass and Zahn, 1996, 2001; Schumacher and Lazarus, 2004; Anderson and Delaney, 2005). Our point is that Late Eocene Tethys closure is a previously unappreciated factor in this global re-organisation.

4. Conclusions

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

Initial Arabia-Eurasia plate collision and closure of the Tethys Ocean provides four complementary mechanisms for reducing atmospheric CO_2 and global cooling: 1) the waning of pre-collision arc magmatism, 2) storage of organic carbon in the

Paratethyan basins, 3) an increase in silicate weathering, 4) re-organisation of ocean currents, consistent with global end Eocene increases in ocean current vigour, organic productivity and hence CO_2 drawdown. We contend that all these mechanisms acted together to help take the Earth across a threshold into the icehouse world at the Oi-1 event.

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Figures

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

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

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

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

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

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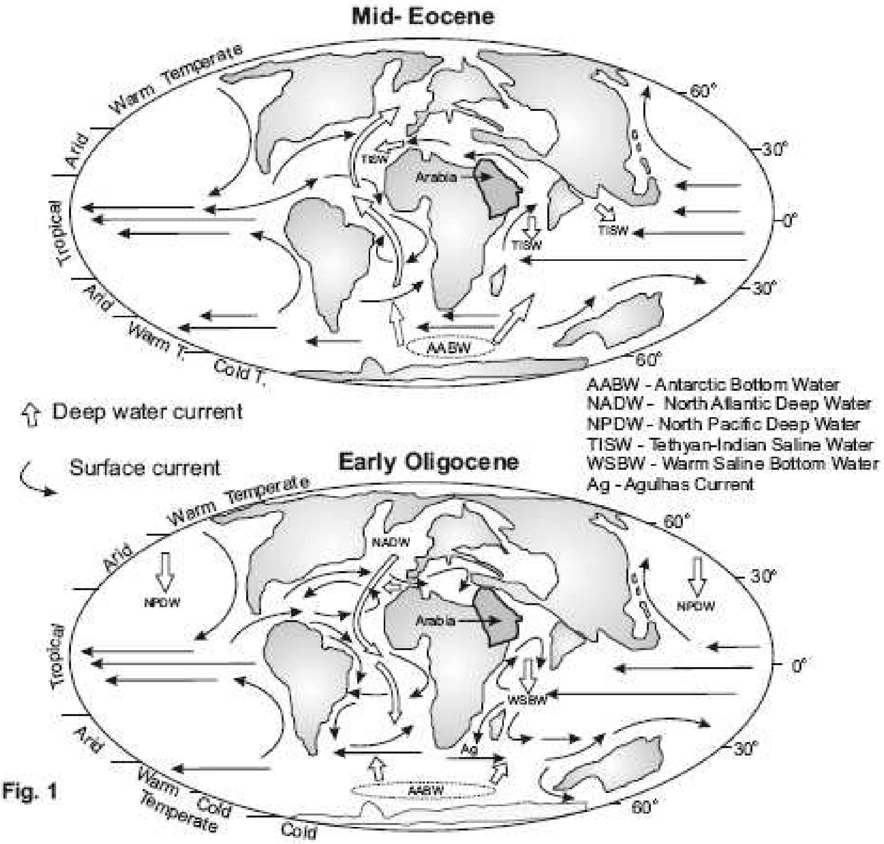
Mid- Eocene
Early Oligocene
Deep water current
Surface current
AABW - Antarctic Bottom Water
NADW - North Atlantic Deep Water
NPDW - North Pacific Deep Water
TISW - Tethyan-Indian Saline Water
WSBW - Warm Saline Bottom Water
Ag - Agulhas Current
Arabia
TISW
AABW
NADW
NPDW
WSBW
Ag
Arid
Warm Temperate
Tropical
Warm T.
Cold T.
Cold
60°
30°
0°
Fig. 1

627

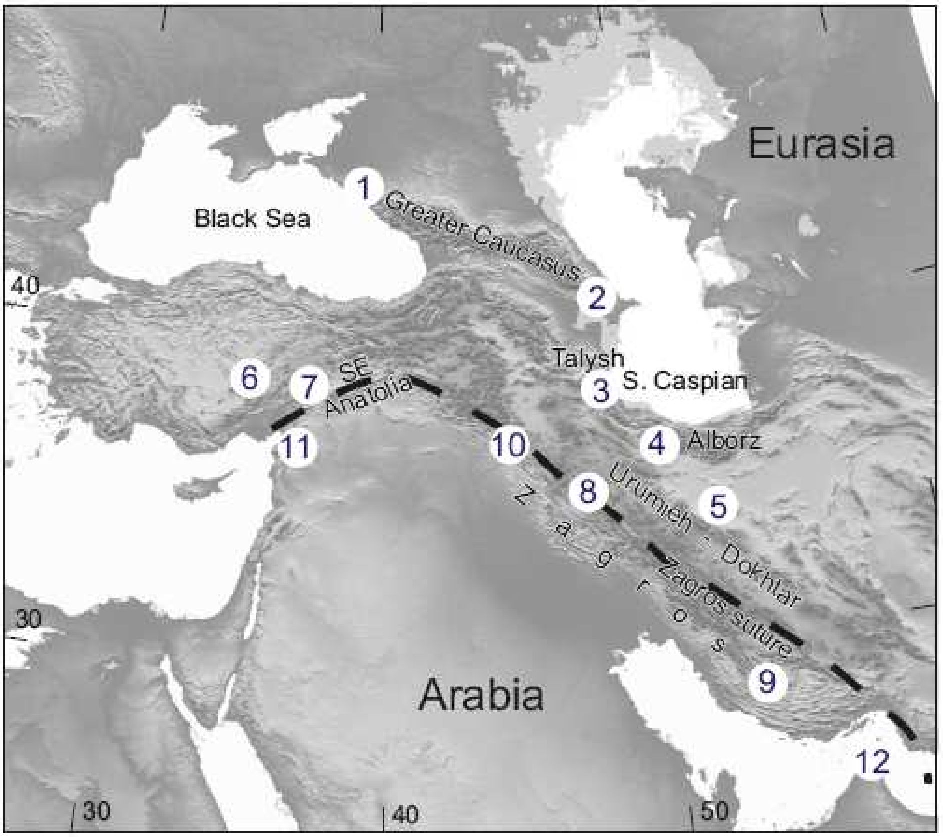
Eurasia
Black Sea
Greater Caucasus
Talysh
S. Caspian
SE Anatolia
Alborz
Urumieh - Dokhtar
Zagros
Zagros suture
Arabia
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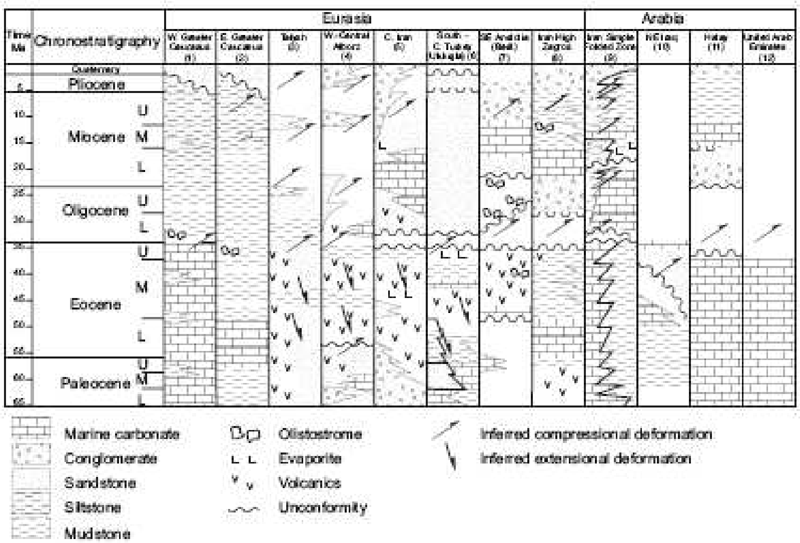
Fig. 2

628


Eurasia
Arabia
Time Ma
Chronostratigraphy
Quaternary
Pliocene
Miocene
Oligocene
Eocene
Paleocene
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M
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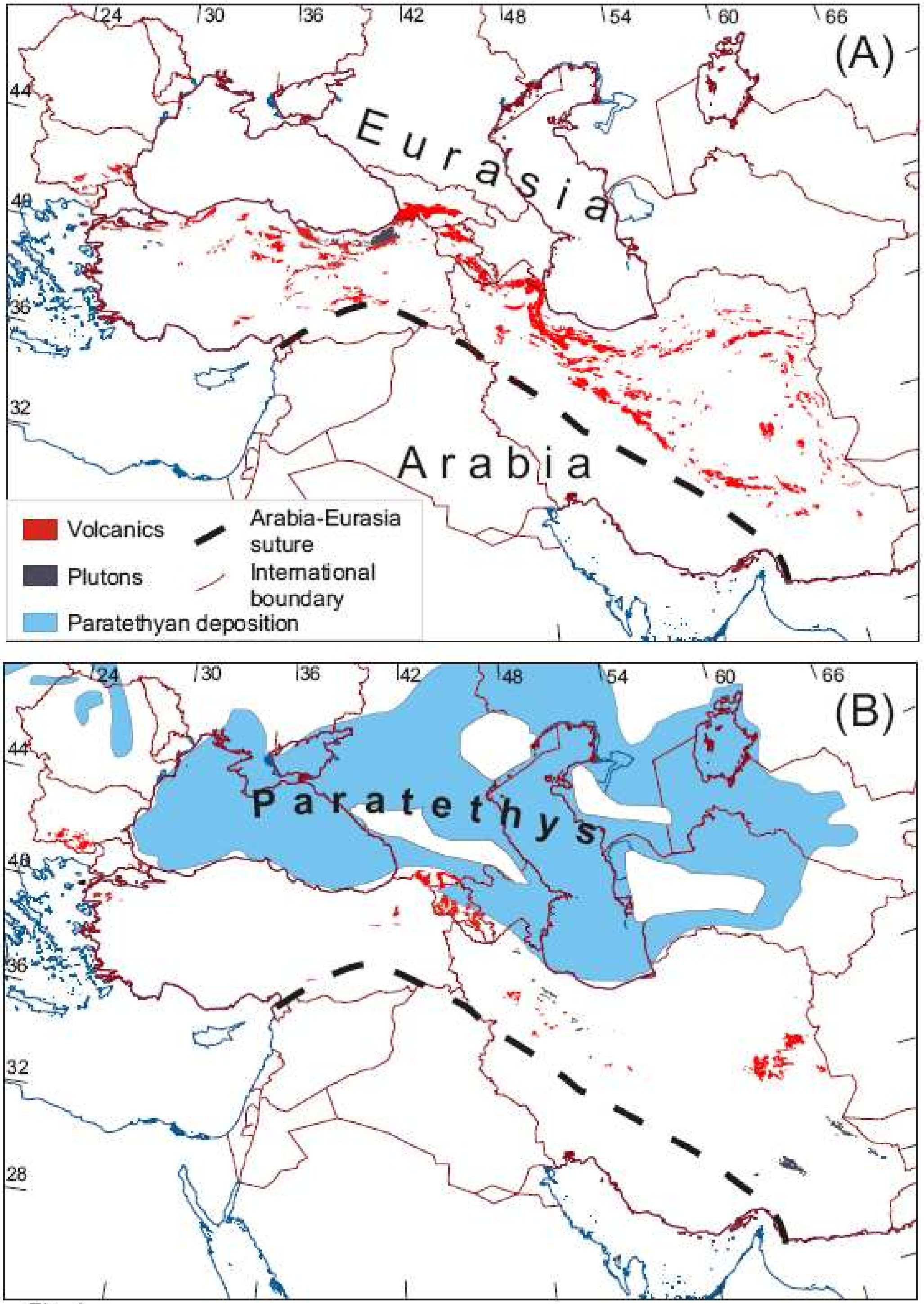
Marine carbonate
Conglomerate
Sandstone
Siltstone
Mudstone
Olistostrome
Evaporite
Volcanics
Unconformity
Inferred compressional deformation
Inferred extensional deformation

Fig. 3

629


24
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(A)
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Eurasia
Arabia
Volcanics
Plutons
Paratethyan deposition
Arabia-Eurasia suture
International boundary
24
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(B)
44
40
36
32
28
Paratethys

630 Fig. 4