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SILJA K. HÜSING1,*, WILLEM-JAN ZACHARIASSE2, DOUWE J. J. VAN HINSBERGEN1, WOUT KRIJGSMA1, MURAT INCEO˘Z3, MATHIAS HARZHAUSER4, OLEG MANDIC4 & ANDREAS KROH4

1Paleomagnetic Laboratory “Fort Hoofddijk”, Department of Earth Sciences, Utrecht University, Budapestlaan 4, 3584 CD Utrecht, The Netherlands
2Stratigraphy and Paleontology Group, Department of Earth Sciences, Utrecht University, The Netherlands
3Department of Geology, Fırat University, Elazığ, Turkey
4Natural History Museum Vienna, Austria
*Corresponding author (e-mail: huesing@geo.uu.nl)

Abstract: The Oligocene–Miocene was a time characterized by major climate changes as well as changing plate configurations. The Middle Miocene Climate Transition (17 to 11 Ma) may even have been triggered by a plate tectonic event: the closure of the eastern Tethys gateway, the marine connection between the Mediterranean and Indian Ocean. To address this idea, we focus on the evolution of Oligocene and Miocene foreland basins in the southernmost part of Turkey, the most likely candidates to have formed this gateway. In addition, we take the geodynamic evolution of the Arabian–Eurasian collision into account.

The Muş and Elazıg basins, located to the north of the Bitlis–Zagros suture zone, were most likely connected during the Oligocene. The deepening of both basins is biostratigraphically dated by us to occur during the Rupelian (Early Oligocene). Deep marine conditions (between 350 and 750 m) prevailed until the Chattian (Late Oligocene), when the basins shoaled rapidly to subtidal/intertidal environment in tropical to subtropical conditions, as indicated by the macrofossil assemblages. We conclude that the emergence of this basin during the Chattian severely restricted the marine connection between an eastern (Indian Ocean) and western (Mediterranean) marine domain. If a connection persisted it was likely located south of the Bitlis–Zagros suture zone. The Kahramanmaras basin, located on the northern Arabian promontory south of the Bitlis–Zagros suture zone, was a foreland basin during the Middle and Late Miocene, possibly linked to the Hatay basin to the west and the Lice basin to the east. Our data indicates that this foreland basin experienced shallow marine conditions during the Langhian, followed by a rapid deepening during Langhian/Serravallian and prevailing deep marine conditions (between 350 and 750 m) until the early Tortonian. We have dated the youngest sediments underneath a subduction-related thrust at c. 11 Ma and suggest that this corresponds to the end of underthrusting in the Kahramanmaras region, i.e. the end of subduction of Arabia. This age coincides in time with the onset of eastern Anatolian volcanism, uplift of the East Anatolian Accretionary Complex, and the onset of the North and East Anatolian Fault Zones accommodating westward escape tectonics of Anatolia. After c. 11 Ma, the foreland basin south of the Bitlis formed not (or no longer) a deep marine connection along the northern margin of Arabia between the Mediterranean Sea and the Indian Ocean. We finally conclude that a causal link between gateway closure and global climate change to a cooler mode, recorded in the Mi3b event (δ18O increase) dated at 13.82 Ma, cannot be supported.

Tectonic closure and opening of marine gateways is suggested to have led to substantial reorganization of surface and deep ocean water currents and may have caused important changes in global climate. The closure of the Panama Isthmus between 3.0 and 2.5 Ma has influenced the Gulf Stream, triggering major Northern Hemisphere glaciations (Bartoli et al. 2005; Schneider & Schmittner 2006). The opening of the Drake Passage allowed the start of the Antarctic Circumpolar Current which might have initiated the abrupt climate cooling around the Eocene/Oligocene boundary and the extensive growth of Antarctic ice sheets (Livermore et al. 2005). The restriction of water exchange across

the former straits between Spain and Morocco resulted in the desiccation of the Mediterranean Sea during its Messinian Salinity Crisis (Hsiü et al. 1973). Likewise, the disconnection of the Indian Ocean and the Atlantic/Mediterranean water masses has been suggested to have caused a major middle Miocene climate change, widely recognized in both the marine (Woodruff & Savin 1989; Flower & Kennett 1994; Zachos et al. 2001; Bicchi et al. 2003) and the terrestrial record (Krijgsman et al. 1994). It is this disconnection that forms the scope of this paper.

The middle Miocene is a period characterized by major environmental changes during which the Earth’s climate gradually progressed into a colder mode (Zachos et al. 2001). The Miocene Climate Optimum between 17 to 15 Ma was followed by an interval of global climate variability between 15 and 14 Ma, marked by atmospheric and oceanic cooling, East Antarctic Ice Sheet growth, and carbon cycle variability (Woodruff & Savin 1989; Flower & Kennett 1994; Zachos et al. 2001). Seven major δ18O shifts, Mi1 to Mi7, to higher (= colder) values documented in marine records of the Atlantic reflect brief periods of increased glaciations (Miller et al. 1991; Wright et al. 1992; Miller et al. 2005). The Mi3a, Mi3b and Mi4 events between about 14.5 and 12.5 Ma represent the middle Miocene δ18O increase, leading the global climate into a colder mode at the same time as the onset of the Antarctic glaciations (van der Zwaan & Gudjonsson 1986; Abels et al. 2005; Miller et al. 2005).

A direct relationship between the Middle Miocene Climate change, whether recorded in oxygen or carbon isotopes, marine or terrestrial fauna, and the closure of the eastern Tethys gateway has so far never been proven, although many studies suggest a causal link between the two events (e.g. Woodruff & Savin 1989; Rögl 1999; Flower & Kennett 1993). Part of the problem is that the sediments that were deposited in the eastern Tethys gateway have on many occasions not been recognized or properly dated. In addition, the chronological sequence of tectonic processes involved in the convergence of the Eurasia and African–Arabian plates is complex and actively debated (see Garfunkel 1998, 2004; Golonka 2004). To assess the timing of gateway closure along the northern Arabian promontory, the major geodynamic processes of the Arabia–Eurasia collision and their tectonic responses have to be taken into account. According to reconstructions of Jolivet & Facenna (2000) and Bellahsen et al. (2003), Arabia collided first in the eastern Anatolian/western Iranian region around 30 Ma ago. Consequently, it gradually rotated counterclockwise leading to diachronous collision eastward from Southeastern Anatolia towards the Persian Gulf (Hessami et al. 2001). Therefore, we decided to study the southernmost flysch deposits in eastern Anatolia (Fig. 1), these being the most likely candidates to represent the youngest sediments deposited just prior to the disconnection of the Indian–Arabian gateway.

**Geodynamic and geological context**

The continental collision of the African–Arabian plate with the Eurasian plate resulted in a tectonic collage in eastern Anatolia that is generally subdivided into: (1) the eastern Rhodope–Pontide Arc in the north; (2) the East Anatolian Accretionary Complex consisting of an ophiolitic mélange overlain by Paleocene to upper Oligocene sediments; and (3) the Bitlis–Pöttürge Massif tectonically overlying the northern part of the Arabian margin (Fig. 1) (Şengör & Yilmaz 1981; Yılmaz 1993; Tüysüz & Erler 1995; Robertson 2000; Şengör et al. 2003; Agard et al. 2005). As north–south shortening continued between the converging Eurasian and Arabian plate, the relatively soft and resistant East Anatolian Accretionary Complex took up most of the initial post-collisional convergent strain by shortening and thickening (Yilmaz et al. 1998). Around 13–11 Ma, eastern Anatolia underwent rapid uplift and was confronted with onset of widespread volcanism (Dewey et al. 1986; Pearce et al. 1990; Keskin 2003; Şengör et al. 2003), which has been associated with detachment of northward dipping subducted lithosphere (Keskin 2003; Facenna et al. 2006; Hafkenscheid et al. 2006). From this moment onward, the ongoing northward motion of Arabia (still continuing today) (McClusky et al. 2000; Reilinger et al. 2006; Allmendinger et al. 2007), and the retreat of the Hellenic subduction zone to the west (Bercchemer 1977; Le Pichon et al. 1982; Jolivet 2001) led to westward tectonic escape of Anatolia along the North and East Anatolian Faults (Dewey & Şengör 1979; Şengör et al. 1985).

The present-day plate boundary of the African and Eurasian plates is determined by the Bitlis–Zagros suture zone (Robertson 2000 and references therein; Westaway 2003). On the Arabian plate, to the south of the suture zone, Eocene and younger (volcano-) sediments are relatively flat lying. North of the Bitlis–Pöttürge zone, Tertiary marine sediments crop out rarely and the geology is dominated by pre-Neogene basement rocks (metamorphic rocks) and Neogene volcanic rocks. The Bitlis–Pöttürge Massif itself is characterized by a stack of nappes originated on the Eurasian side of the Neotethys (Robertson 2000; Robertson et al. 2004).

The Bitlis–Pöttürge Massif runs from southeastern Turkey to the eastern Mediterranean basin into the Cyprus arc, where it meets the East Anatolian
Fault (EAF). Here the structure becomes more complex with several sub-parallel southwestwards running faults and thrusts. The East Anatolian Fault is a 2–3 km wide, active left-lateral strike-slip fault extending from Antakya in the west to Karliova in the NE, where it meets the eastern termination of the North Anatolian Fault (NAF) (Figs 1 & 2; EAFZ and NAFZ). The NAF is a right-lateral strike-slip fault extending over a length of about 1300 km westward. The relative Africa–Arabia motion is taken up by strike-slip displacement along the Dead Sea Fault (Jolivet & Faccenna 2000), while the Africa–Anatolia motion is taken up by subduction south of Cyprus. The overall convergence between Arabia and Anatolia is taken up along the North and East Anatolian fault zones (NAFZ and EAFZ) (Fig. 2) (e.g. McClusky et al. 2000; Şengör et al. 2005). There is general consensus that the NAFZ and EAFZ had the majority of their displacement in Plio-Pleistocene times (Barka 1992; Westaway 2003, 2004; Hubert-Ferrari et al. 2008) although incipient motion may have been as early as late Serravallian/early Tortonian (c. 12 to 11 Ma) (Dewey et al. 1986; Hubert-Ferrari et al. 2002, 2008; Bozkurt 2003; Şengör et al. 2005).

The region that comprises the eastern Tethys gateway has thus been subjected to plate convergence and subduction. Şengör et al. (2003) suggested that this subduction led to southward migrating accretion of nappes and overlying deep-marine foreland basin deposits, even though individual basins that may reflect such evolution have not been identified in the geological record, which is, at least in part, due to the young volcanic sequences covering a large part of eastern Turkey. If southward accretion of nappes indeed occurred, one should be able to identify southward younging flysch deposits (e.g. van Hinsbergen et al. 2005a).

A foredeep likely remains present until continent–continent collision and subsequent slab break-off stalls convergence and the collision zone is uplifted. Even though small marine basins may remain, the long distance between the Persian Gulf and the Mediterranean Sea makes foredeeps the most promising basins to have formed the gateway between these water masses. In the following paragraphs we will present and discuss the evolution of foredeep basins in SE Turkey in the light of the closure of the eastern Tethys gateway.

**Basin evolution**

The Arabian foreland is separated from the East Anatolian Accretionary Complex (EAAC) by the Bitlis–Pötürge Massif (Fig. 1). The area of this massif corresponds to the compression zone located...
Fig. 2. Outline of schematic geological map of SE Anatolia in Southeastern Turkey with major tectonic structures. Note the three boxes indicating the studied areas: Muş in the easternmost part and Elazığ, both north of the Bitlis–Zagros Suture zone and Kahramanmaraş south of the Bitlis–Zagros Suture Zone (drawn after Geological map of Turkey (Şenel 2002)).
between the two continental crusts, Eurasia and African–Arabian. The massif was stacked to form a nappe complex during the closure of the Neo-Tethys by the middle Miocene (Dewey 1986 and references herein).

We have studied the southernmost flysch deposits in the eastern Anatolian orogenic system. These are found in the Muş and Elazıg basins, both north of the Bitlis–Pötürge Massif, and the Kahramanmaraş basin located south of the Bitlis–Pötürge Massif and near the triple junction of the Arabian, Eurasian and Anatolian plates (Fig. 2).

Geological setting of the Muş basin

The Muş basin is an elongated structure located north of the Bitlis–Pötürge Massif and east of the North and East Anatolian Fault (Figs 2 & 3). According to previous studies (Şaroğlu & Yılmaz 1986; Sancay et al. 2006) the basin contains upper Eocene to lower Miocene limestones, marls and turbiditic sandstones with marine sedimentation continuous from the Oligocene to Aquitanian. These deposits overlay an upper Cretaceous ophiolitic mélangé. Şaroğlu & Yılmaz (1986) suggested that lower Miocene limestones are widespread in the northern part of the Muş area, while middle Miocene strata were not found. These sequences are unconformably covered by allegedly upper Miocene and younger continental clastics and volcanics (Şaroğlu & Yılmaz 1986; Sancay et al. 2006). Detailed biostratigraphy was carried out mainly based on dinoflagellates and palynomorphs yielding a Rupelian (early Oligocene) to Aquitanian (early Miocene) age (Sancay et al. 2006). The occurrence of the benthic foraminiferal family of Miogypsinidae was interpreted as possible indicator for a connection with the Indo-Pacific during the Oligocene (Sancay et al. 2006).

We sampled two sections in the Muş basin (Fig. 3). The eastern transect comprises allegedly Eocene–Oligocene clastics in the northern part of the basin, and Oligocene flysch sediments followed by marine limestones which are covered by volcanics. The second transect in the western part of the basin covers the transition from marls to limestones, assuming it is equivalent to the uppermost part of the eastern succession. The entire succession gently dips towards the NW.

The base of the eastern section (east transect in Fig. 3) is determined by a thrust zone emplacing allegedly Eocene clastic sediments onto Pliocene deposits (see geological map of Turkey, Şenel 2002) (Fig. 3). The first 20 m of the studied section is characterized by an alternation of conglomerates, clays, sands, and silts (Fig. 4). A layer of limestone (1.5 m) with shell fragments and the presence of large gastropods clearly indicate shallow...
Fig. 4. Lithological column of the studied sections in the Mus basin with the biostratigraphic results. The age model is based on planktonic foraminifer occurrences and the macrofossil assemblage in the uppermost 40 m (mainly limestones and sands) of the stratigraphy. Planktonic foraminifer occurrences have been correlated to planktonic foraminifer zones, which, in turn, are tied to stages during the Oligocene leading to a correlation to the Geological Time Scale. See legend for key to lithologies, structures and fossils.
marine conditions. This sequence is followed by a thick (about 1.3 km) succession of alternating clay and sandstone. Occasionally conglomeratic layers, characterized by angular, unsorted material, occur in the succession. These layers have thicknesses of up to 10 m and are interpreted as debris flows. The sandstone layers show typical transport characteristics such as fining upwards, Bouma sequence, flute casts and fossil fragments indicating a turbiditic origin. These turbidites occur as massive sandstone layers of thicknesses of up to 15 m or as several thinner (up to 50 cm) turbidite layers, probably representing individual events. Only minor slumping, indicating an unstable submarine paleoslope, and folding occur throughout the succession. The upper part of the section shows shoaling characterized by shallow marine limestone, containing echinoderms, bivalves and gastropods, followed by continental clastics.

The western transect (west transect in Fig. 3) is dominated by bluish clay with occasional red sediments. This is followed by a thick, about 100 m, sequence of alternating softer bluish sands, brownish sands and indurate bluish sands, probably all of marine origin. These sediments are overlain by coral limestones, which, in turn, are covered by volcanic rocks, probably of Miocene age. This succession also clearly indicates shoaling towards the top.

**Biostratigraphic results of the Muş basin**

For biostratigraphy, samples were collected at about every 20 m from both the western and eastern transect (Fig. 3). Not every sample proved to be useful for biostratigraphy or paleobathymetry. The number of foraminifers is extremely variable and most likely fluctuate in pace with changes in terrigenous clastic input. Preservation is generally poor with specimens mostly recrystallized and frequently distorted. Samples from the upper 300 m of the western section are barren in planktonic foraminifers.

The low diversity in planktonic foraminiferal fauna in both sections is dominated by globoquadrids and catapsydracids with occasional occurrences of Globigerina ciperoensis and Globigerina angulisuturalis and clearly points to an Oligocene age for the eastern and western lower western section (Fig. 4) (Berger & Miller 1988; Spezzaferri & Premoli Silva 1991).

The basal part of the eastern section correlates to planktonic foraminiferal biozone P19 of Berger & Miller (1988) on the basis of trace occurrences of specimens identical to Turborotalia ampliapertura. This biozone is Rupelian (early Oligocene) in age (Fig. 4) (Berggren et al. 1995). The lowermost occurrence of Globigerina angulisuturalis is recorded at 950 m (TR 221) in the eastern section which together with the highest occurrence of Paragloborotalia opima opima (at 1045 m (TR222)) indicates that the middle part of the Muş section correlates to planktonic foraminiferal biozone P21 of Blow (1969) and Berger & Miller (1988) which is latest Rupelian to early Chattian in age (Berggren et al. 1995). The absence of Paragloborotalia opima opima from sample level 1045 m (TR 222) upward in the eastern transect and the occurrence of typical Paragloborotalia pseudokugleri and even of forms transitional between Paragloborotalia pseudokugleri and Paragloborotalia kugleri at the top of the section (1360 m (TR 232)) indicate that the upper part extends upwards into the lower part of planktonic foraminiferal biozone P22 of Berger & Miller (1988) being Chattian in age (Berggren et al. 1995). This is confirmed by the presence of Paragloborotalia siakensis and Globigerinoides primordius in the youngest samples. Both species make their first appearance in the lower part of biozone P22 together with Paragloborotalia pseudokugleri (Berger & Miller 1988; Spezzaferri 1994).

In the western section, the co-occurrence of Globigerina angulisuturalis and Paragloborotalia opima opima at 2 m and 195 m (TR 202 and TR 210) indicates that the lower 200 m correlates to the interval between 900 and 1100 m in the eastern section. Both these intervals belong to biozone P21. This interval is followed by sediments that are barren in planktonic foraminifers but relatively rich in shallow water benthic foraminifers.

The macrofossil assemblage of the uppermost 40 m in the eastern transect comprises bivalves, gastropods and echinoids. The assemblage is diminished by complete aragonite leaching. Nevertheless, the fauna is age indicative and allows palaeoecological interpretations. The mollusc fauna comprises typical Oligocene taxa such as the gastropod Ampullinospis cressatinus (Lamarck 1804) and the bivalves Amussiopecten labadyei (d'Archiac & Haime 1853) and Ringicardium bukkhiam (Telegdi-Roth 1914). Some species such as Dilatilabrum sublatissinum (d'Orbigny 1852), Strombus cf. praecebedens Schaffner 1912, Cordiopsis incrassatus (Nyst 1836), Amussiopecten subpleuronectes (d'Orbigny 1852), and Hyotissa hyotis (Linnaeus 1758) appear during the Chattian and persist into the Miocene.

An important biostratigraphic feature is the co-occurrence of the pectinids Amussiopecten labadyei and A. subpleuronectes and the occurrence of transitional morphs. This evolutionary phase is recorded so far only from the upper Chattian (Mandic 2000). Especially in the Iranian Qom Basin, this assemblage co-occurs with the larger foraminifera Eulepidina dilatata. The last occurrence of Amussiopecten labadyei precedes the first occurrence of Miogypsinoidea which roughly coincides
with the base of the early Miocene. The entire mollusc assemblage is therefore pointing to a late Chattian age. This dating is supported by the echinoid fauna. *Parascutella subrotundaeformis* (Schauroth 1865), a sand dollar which occurs most commonly in Northern Italy, is restricted to the Chattian and Aquitanian.

Comparable assemblages are described from the upper Chattian of the central Iranian Qom Formation (Mandic 2000; Harzhauser 2004; Reuter et al. 2007) and along the entire northern coast of the Western Tethys (Harzhauser et al. 2002). A relation to the Central Paratethys is indicated by the occurrence of *Ringicardium bukkianum*, which is known from the Lower Egerian (Upper Chattian) deposits of Hungary (Baládi 1973). The faunistic relations towards the east are low. Only *Dilatilabrum sublatissimus* (d’Orbigny 1852) reaches to the Zagros Basin and the Arabian shelf during the Aquitanian (Harzhauser et al. 2007). The echinoderm *Clupeaster waageni* (Duncan & Sladen 1883), in contrast, represents ties with the echinoid fauna of the Lower Indus Basin.

Numerical ages for the basin fill are provided by three planktonic foraminiferal bioevents. However, equating highest and lowest occurrences (ho and lo) with the Last Occurrence (LO) and First Occurrence (FO) of these species should be accepted with reservation because the positions are poorly delineated due to large sampling distances in combination with scarcity and poor preservation of the age diagnostic species.

The oldest bioevent in the Müş section is the lowest occurrence of *Turborotalia ampniapertura* some 300 m above the base of the eastern section (TR 190). The LO of this species is calibrated at 30.3 Ma (Berggren et al. 1995) providing a minimum age for the base of the Müş section. The age for the top of the eastern section should be slightly younger than the age of 25.9 Ma for the FO of *Paragloborotalia pseudokugleri* (Berggren et al. 1995) because of the presence of paragloborotalids being transitional between *Paragloborotalia pseudokugleri* and *Paragloborotalia kugleri*. The ho of *Paragloborotalia opima opima* at 1045 m (TR 222) in the eastern section provides an extra age calibration point of 27.456 Ma being the calibrated age for the LO of *Paragloborotalia opima opima* at ODP Site 1218 (Wade et al. 2007). The dating of the top of the section is in accordance with the macrofauna which strongly indicates a late Chattian age for the upper 40 m of the eastern transect.

No numerical ages are provided for the western section. However, based on the co-occurrence of *Globigerina angulisuturalis* and *Paragloborotalia opima opima* in the lower 200 m, this interval correlates to the biozone P21. The upper 300 m lack any age diagnostic planktonic foraminifer.

**Palaeoenvironmental interpretations for the Müş basin**

Benthic foraminifers in the sections were further more used to estimate the depositional depth. The commonly used method of calculating depth by determining the ratio between planktonic and benthic foraminifers (van der Zwaan et al. 1990; van Hinsbergen et al. 2005b) is not reliable here due to significant downslope transport (seen in presence of notorious epifytes and shallow water benthic foraminifers such as *Pararotalia* and *Amphistegina*) and poor preservation. Instead, we focus on the deepest water benthic foraminiferal depth markers (for list see van Hinsbergen et al. 2005b) and the macrofossils. In the eastern section, the depositional environment of the lower 20 m is characterized by shallow marine conditions, indicated by shell fragments in the limestone. However, a rapid deepening trend occurs at about 50 m indicated by the presence of benthic foraminiferal depth markers (typically *Cibicides* (pseudo) *ungerianus*, *Gyroidea* spp. *Uvigerina* spp. and occasionally *Oridorsalis* spp.), and the absence of markers for deeper water, which points at a depositional depth range of 350 to 750 m (the upper limit is constraint by the occurrence of *Oridorsalis* spp. after van Hinsbergen et al. 2005b). Towards the top of the eastern section rapid shoaling is evident from the presence of macrofossils. Both the molluscs and echinoderms of the uppermost 40 m indicate a shallow marine, tropical to subtropical, depositional environment with sand bottoms and algal or sea grass patches. Giant conchs such as *Dilatilabrum sublatissimus* (d’Orbigny 1852) are found today in sea grass meadows and sheltered lagoons, where they live partly buried in the soft substrate (Bandel & Wedler 1987). Similarly, the extant representatives of the oyster *Hyotissa hyotis* prefer shallow subtidal habitats where they are attached to rocks and corals (Slack-Smith 1998). Extant *Echinolampas* and *Clupeaster*, too, occur most commonly on sandy sediments with sea grass patches (Hendler et al. 1995).

In the western section a shoaling trend in the upper 250 m is observed by the relatively rich occurrence of shallow water benthic foraminifers and occasional red sediments. The differences between west and east suggest that the western part of the Müş basin shoaled more rapidly or earlier during the Chattian than the eastern part.

**Implications for the Müş basin**

Based on the occurrence of turbidites, slumping and minor folding, this about 1.5 km thick marine succession is interpreted as deposits of a deep marine basin.
Shallow marine conditions during the Rupelian (P19) were replaced by rapid deepening of the basin during biozone P22, late Chattian. The end of the flysch deposition during the Chattian marks the emergence of the basin which probably remained shallow marine until the late Chattian. Considering the biostratigraphic ages, a sedimentation rate between 15 and 27 cm/ka is calculated. The constant water depth of 350 to 750 m during deposition indicates approximately 2 km of subsidence throughout the Oligocene, followed by rapid uplift and exposure of the succession after the late Chattian. Our biostratigraphic dates based on planktonic foraminifers in the flysch deposits corroborate the ages published previously based on dinoflagellates and palynomorphs (Sancay et al. 2006).

Geological setting of the Elazıg basin

The studied Gevla section is situated in the easternmost part of the Elazıg basin, about 40 km NE of Elazıg (Fig. 2). The basin has been studied by several workers; however, the literature has been published mostly in Turkish (see Aksoy et al. 2005) and no detailed information is available for the easternmost part of the basin. At present, the basin fill is exposed in an NE–SW belt in the eastern Taurides of Anatolia. The generalized stratigraphy of the Tertiary sediments has been described as: lower Paleocene continental deposits at the base, followed by upper Paleocene to lower Miocene marine deposits and finally Pliocene to Quaternary continental deposits. The basement of the Elazıg basin is formed by Permo-Triassic metamorphic rocks, namely Keban Metamorphics, which were emplaced over upper Cretaceous magmatic rocks north of Elazıg (Perinçek 1979; Perinçek & Özkaya 1981; Aktaş & Robertson 1984; Bingöl 1984; Aksoy et al. 2005).

Detailed stratigraphic, sedimentological and tectonic characteristics of the Elazıg area have been discussed elsewhere (e.g. Perinçek 1979; Perinçek & Özkaya 1981; Aktaş & Robertson 1984; Bingöl 1984; Cronin et al. 2000a, b; Aksoy et al. 2005). From the late Paleocene, shallow marine carbonates, deposited in an extensional back-arc setting, were accumulated when the basin further subsided until Middle to Late Eocene (Aksoy et al. 2005). During Oligocene to early Miocene, after reaching its maximum extend during the Middle to Late Eocene, deposition was restricted to the N–NW and became progressively shallower, indicated by Oligocene reeval limestones until the final subaerial exposure at the end of Oligocene. Marine Miocene deposits are restricted to small areas in the basin and more widespread north of the basin. From Middle Miocene onwards the basin was affected by a strong, north–south, compression. Later, Pliocene to Pleistocene alluvial fan, fluvial and lacustrine sediments were deposited covering Early Miocene sediments (Cronin et al. 2000a, b; Aksoy et al. 2005).

In this setting, we studied a section situated in the easternmost part of the Elazıg basin. According to the geological map of Turkey (Şenel 2002), in the area east of the town Basyurt (Fig. 5), Lower to Middle Eocene continental clastics unconformably overly Mesozoic ophiolitic mélangé. These clastic sediments are, in turn, overlain by either Miocene–Pliocene clastic or volcanic rocks.

The basal part of the studied Gevla succession, about 15 km NE of Basyurt, starts with bluish marine clay containing bivalves, followed by an alternation of clay and sandstone (the sandstones are up to 50 cm thick or about 5 m thick with cross bedding) (Fig. 6). A distinct layer with abundant bivalves and gastropods is located at about 50 m stratigraphic position. Three distinct limestone layers occur between about 100 m and 260 m stratigraphic level. The first one, at about 100 m, is a nodular limestone with shell fragments, sponges (up to 30 cm) and corals, followed by two nummulitic limestone horizons, at 244 m and 255 m. This is followed by about 400 m of blue clay grading into a 600 m thick succession of alternating clay and sandstone, whereby the sand layers show typical transport characteristics, such as shell fragments,
displaced nummulites and gastropods (for instance at 663 m and 1158 m), fining upward sequences and cross bedding. These layers are interpreted as turbiditic in origin. This succession is followed by about 300 m of blue clay, and the section ends with a 50 m thick limestone with bivalves (up to 5 cm), and clayey intervals with well preserved echinoderms, sponges and corals. These limestones, in turn, are covered by Miocene volcanic rocks. In total, the section is about 1.6 km thick.

Slumping at several levels within the succession indicates an unstable submarine slope. Internal folding is not observed within the succession. The entire succession gently dips towards the NW.

Biostratigraphic results of the Elazığ basin

Hand samples were collected from about every 20 m throughout the entire section, but not every sample contained (diagnostic) planktonic and/or benthic foraminifers. The number of foraminifers is extremely variable and most likely fluctuates with changes in terrigenous clastic input. Preservation is generally poor, with specimens mostly recrystallized and frequently distorted. The overall aspects of the planktonic foraminiferal fauna in this section is similar to that of the Muş section, which means that the foraminiferal fauna is dominated by globoquadrinids and catapsydracids with occasional occurrences of *Globigerina ciperoensis* and *Globigerina angulisuturalis* pointing to an Oligocene age for this section (Fig. 6) (Berger & Miller 1988; Spezzaferri & Premoli Silva 1991). The presence of *Turborotalia ampliapertura* up to and including level 287 m (TR 244) provides evidence that the lower part of the section correlates with planktonic foraminiferal biozone P19 of Berger & Miller (1988), which is late Rupelian, early Oligocene, in age (Berggren et al. 1995). The lowest occurrence of *Globigerina angulisuturalis* is recorded at 477 m (TR 250) which in terms of the zonal scheme of Berger & Miller (1988) would mark the top of biozone P20 although it should be noted that *Globigerina angulisuturalis* is neither frequent in this section nor does it display very prominent U-shaped sutures. Typical *Paragloborotalia opima opima* is present from level 317 m (TR 245) up to and including level 1445 m (TR 293) indicating that the larger part of the Gevla section correlates with planktonic foraminiferal biozone P19 of Berger & Miller (1988), which is late Rupelian, early Oligocene, in age (Berggren et al. 1995). The lowest occurrence of *Globigerina angulisuturalis* is recorded at 477 m (TR 250) which in terms of the zonal scheme of Berger & Miller (1988) would mark the top of biozone P20 although it should be noted that *Globigerina angulisuturalis* is neither frequent in this section nor does it display very prominent U-shaped sutures. Typical *Paragloborotalia opima opima* is present from level 317 m (TR 245) up to and including level 1445 m (TR 293) indicating that the larger part of the Gevla section correlates with planktonic foraminiferal biozone P19 of Berger & Miller (1988), which is late Rupelian, early Oligocene, in age (Berggren et al. 1995).
P21 of Blow (1969) and Berggren et al. (1995) which in terms of chronostратigraphy is latest Rupelian to early Chattian in age (Berggren et al. 1995). The top of the section post-dates the highest occurrence of Paragloborotalia opima opima, and correlates to the basal part of the late Chattian planktonic foraminiferal biozone P22 (Berger & Miller 1988), which is evidenced by the joint presence of Paragloborotalia opima opima, Globigerina ciperoensis, Globigerina angulisuturalis and Globigerinoides primordius.

The macrofossil assemblage from the upper 50 m in this section is similar to the assemblage in the uppermost 40 m of the eastern transect in the Muş basin (Figs 3 & 4). Both assemblages bear a typical late Chattian pectinid fauna with Amasistopecten labadyei and A. subpleuronectes. Pecten arcuatus (Brocchi 1814), a widespread Oligocene species, a typical Western Tethys element, is present as well, along with the thin-shelled lucinid bivalve Anodontia globulosa (Deshayes, 1830). The dominance of such thin shelled species might indicate a slightly deeper environment than in the corresponding section of the Muş basin, yet not deeper than the medium deep sublittoral environment (Mandic and Piller 2001).

The LO of Turborotalia ampliapertura has been calibrated to 30.3 Ma within Chron 11r (Berggren et al. 1995). The highest occurrence of this species in this level 287 m (TR 244) therefore suggests an age older than 30.3 Ma for the bottom of the section. The LO of Paragloborotalia opima opima in level 1445 m (TR 293) has been recently recalibrated to 27.456 Ma within Chron 9n at ODP Site 1218 (Wade et al. 2007). This age provides a maximum age for the top of the section since the highest occurrence of Paragloborotalia opima opima occurs near the top of the section (TR 293). A correlation of the upper 50 m of the section to biozone P22 is supported by the mollusc fauna which indicates a late Chattian age.

Paleoenvironmental interpretations
for the Elazığ basin

The depositional environment during the lower Rupelian (biozone P19) was first shallow marine as indicated by the occurrence of limestone with corals, bivalves and gastropods. However, the depositional environment rapidly deepened as indicated by benthic foraminiferal depth marker species (Cibicides (pseudo-)ungerianus, Gyroidina spp. Uvigerina spp. and occasionally Oridorsalis spp.). Their presence up to the top indicates that the basin was 350 to 700 m deep during much of the Oligocene. The benthic foraminifers do not give any evidence for shoaling, although the limestone deposits at the top of the section and their macrofossils indicates a medium to shallow subtidal environment for the late Chattian.

Implications for the Elazığ basin

The first 260 m of the studied section was deposited under shallow marine conditions during Rupelian (biozone P19). This was followed by a rapid deepening during the Rupelian and the deposition of about 1.3 km in a relatively deep marine (300 to 750 m) environment. During the late Chattian (biozone P22), the basin experienced rapid shoaling to medium deep sublittoral conditions, preferred conditions for echinoids and bivalves. The inferred late Chattian age of the macrofossils in the top of the section indicates that the final emergence of the basin must have occurred shortly after the Chattian followed by widespread Miocene volcanism.

The numerous internal slumping and sandstone layers, referred to as turbidites, indicate a submarine, unstable slope. The entire succession is interpreted as flysch deposited in a deep marine basin, comparable to the Muş basin. Thus, during the Oligocene, rapid (15–30 cm/ka) sedimentation of clay and turbidites dominated the basin evolution.

These new biostratigraphic ages differ significantly from the geological map of Turkey (Şenel 2002) where these sediments are indicated as Lower to Middle Eocene. Our data suggests instead that these sediments were deposited under deep marine conditions during the Oligocene, from the Rupelian until the late Chattian, and, additionally, the shallow marine limestones at the top of the section are late Chattian in age. This data also differs from previous studies in the area (e.g. see Aksoy et al. 2005 for a compilation of data from the Elazığ basin) where the Eocene time has been identified as the main period of deep marine deposition and in the Oligocene time shallow marine deposits were restricted to the NW of the Elazığ basin. Our data however indicates that at least the eastern part of the Elazığ basin was deep marine throughout the Oligocene and shoaled and emerged only in the late Chattian, latest Oligocene.

Geological setting of the Kahramanmaraş basin

The Kahramanmaraş basin is located near the triple junction of the Arabian, African and Anatolian plates. As a result of the collision of Arabia and Eurasia along the Bitlis Suture a trough formed in front of the thrust sheets and was consequently filled by thick alluvial sediments and thick turbiditic flysch sequences (Lice Formation) (Şengör & Yılmaz 1981; Perinçek & Kozlu 1983; Karığ & Kozlu 1990;
Yılmaz 1993). According to several studies (Perinc¸ek 1979; Perinc¸ek & Kozlu 1983), the Kahramanmaraş basin was part of this elongated foreland basin extending from Hakkari in southeastern Turkey, close to the border to Iran and Iraq, to Adana in southern Turkey (Fig. 2). This basin was also called the Lice trough (Dewey et al. 1986; Karig & Kozlu 1990; Derman & Atalik 1993; Derman 1999). Eocene deposits in the Kahramanmaraş area are part of the Arabian Platform (Robertson et al. 2004). They indicate a shallow marine depositional environment with local terrestrial input followed by allegedly lower to middle Miocene reefal limestone and claystone (Gül et al. 2005). Oligocene bioclastic limestones are reported only from the margin of the Kahramanmaraş area (Fig. 7) (Karig & Kozlu 1990). Basal shallow marine red-bed and basalt sequences of the Kalecik Formation have an inferred age of late Burdigalian to Langhian (Karig & Kozlu 1990). The retreat of marine conditions and basin deformation was assumed to have taken place in the late Miocene, although the age control was not documented (Karig & Kozlu 1990).

Three separate sections (Figs 7, 8a & b), all north of the city of Kahramanmaraş, are being studied by us. The lower 200 m were sampled in the hills in the southern part of the main basin (Hill section), the following about 4.6 km along the road north of Kahramanmaraş (Road section) and the upper 1.5 km stratigraphic transect near the village of Avcılar (Avcılar section).

The base of the Hill section consists of nummulitic limestones according to the Geological Map of Turkey (Şenel 2002) of Eocene age, followed by red, conglomeratic sediments with several basalt layers.

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**Fig. 7.** Simplified tectonic and geological map of the Kahramanmaraş area including the trajectories of the three studied sections north of the city of Kahramanmaraş: (1) the lowermost 200 m in the Hill section, (2) about 4.6 km of succession along the road (Road section), (3) the upper 1.6 km up to the thrust studied in the Avcılar section in the northernmost part of the region. Refer to Legend for key to lithologies and/or ages (drawn after Geological map of Turkey (Şenel 2002)).
Fig. 8. (a) Lithological column of the Hill section and part of the Road section (c. 3 km) in the Kahramanmaras basin with the biostratigraphic results. The age model is based on planktonic foraminifer occurrences, which delineate the correlation of the Hill section to the Langhian and the first c. 3 km of the Road section to the Serravallian.
Fig. 8. (Continued) (b) Lithological column of the upper c. 1.6 km of the Road section and the Avcılar section in the Kahramanmaraş basin with the biostratigraphic results. The age model is based on planktonic foraminifer occurrences, which delineate the correlation of the upper c. 1.6 km of the Road section to the Serravallian and the lower part of the Avcılar section to the Serravallian, probably overlapping with the Road section, and from c. 700 m to the Tortonian based on the LCO Oof Globigerinoides subquadratus. Refer to Legend in Figure 4 for key to lithologies, structures and fossils. KM = Kahramanmaraş.
The studied section begins with a 200 m thick succession of nodular limestone (10–15 m thick) alternating with bluish marls. The limestones contain macrofossils such as corals, sponges, echinoderms, bivalves and gastropods, indicating a shallow marine environment. This succession grades into an alternation of marl and sandstone layers, which show typical Bouma sequences and flute casts, which are indicative for a turbiditic origin.

The base of the Road section, however, exposes a strongly different succession, where almost 1 km of the stratigraphy is dominated by large conglomeratic lenses. This thick succession of conglomerates contains sand lenses showing cross bedding, indicative of interfingerung of braided river channels. This thick fluvial succession probably forms, at least in part, the lateral equivalent of the shallow marine succession in the Hill section. This conglomeratic succession is followed by a level rich in oysters, indicating a transition into shallow marine conditions. This then grades quickly into a very thick succession of alternating marl and sandstone (as mentioned above) with occasional conglomeratic layers, interpreted as debris flows, which cut through the stratigraphy. The sandstones show typical characteristics for turbidites, such as flute casts and Bouma sequences. Some intervals are dominated by massive sandstone layers and/or debris flows, while others are characterized by mainly clay. Slumping can be easily differentiated from (minor) internal folding, both occurring throughout the section. Internal folding, however, does not occur often. The Road section ends at the highest point in the topography along the road going NNW from Kahramanmaraş and since the stratigraphy dips to the NW, a continuation of the stratigraphy was found to the NE, the Avcılar section (Fig. 7). This section was sampled assuming sufficient overlap with the Road section, until the stratigraphy was cut unconformably by carbonates (Figs 8a & b). The stratigraphy of the Avcılar section also consists of a thick succession (about 1.5 km) of mudstone and sandstone with occasional conglomeratic layers, interpreted as debris flows. No shoaling trend based on sedimentological characteristics has been observed towards the top of the section, which ends abruptly with the overthrusting of pre-Neogene carbonates, which were emplaced roughly from North. The upper 400 m were not exposed, except for a few meters just underneath the thrust (Fig. 8b).

Biostratigraphic results of the Kahramanmaraş basin

About every 20 m hand samples were taken for biostratigraphy. Only few samples, however, turned out to be useful for biostratigraphy and/or paleobathymetry. Benthic foraminifers in the lower part of the Kahramanmaraş basin, from the Hill section (Fig. 8a) are dominated by mioloids and representatives of Ammonia, Textularia, Nonion and Elphidium indicating shallow marine (inner shelf) conditions although some samples, at 146, 182 and 198 m (TR 9, 12 and 14), respectively, contain few planktonic foraminifers such as Globigerinoides trilobus, Globigerinoides obliquus and Orbulina. Their presence would indicate that the lower part of the Kahramanmaraş sequence post-dates the Orbulina datum at 14.74 Ma (Lourens et al. 2004). This age assignment is further constrained by the presence of the calcareous nannofossil Cyclicargolithus abisectus and rare Spenolithus heteromorphus along with the absence of Helicosphaera ampliaperta. This assemblage is tentatively assigned to NN5 (Martini 1971) indicating a Langhian age.

Orbulina is common in the samples from the Road section. The rare occurrences of Globoratalia partimlabiata in the Road section at 165, 175, 2725, 2788 and 2875 m (TR 20, 21, 71, 73 and 77), respectively, are remarkable because they represent the first recording of this species in Turkey. It has been first described from the middle Miocene of Sicily (Ruggieri & Sprovieri 1970) and since then reported from the Mediterranean (amongst others Foresi et al. 1998 and references herein; Hilgen et al. 2000; Turco et al. 2001; Foresi et al. 2002a, b; Hilgen et al. 2003; Abels et al. 2005) and adjacent North Atlantic (Chamley et al. 1986) and even from the Indian Ocean off NW Australia (Zachariasse 1992). Ages of FO and LO of Globoratalia partimlabiata in the Mediterranean have recently been recalibrated at 12.771 and 9.934 Ma (Hüsing et al. 2007). Its presence in the basal part of the Road section along with Globigerinoides subquadratus at 4224 m (TR 114) indicates that the larger part of the Road section, up to 4200 m, is Serravallian, Middle Miocene, in age, since the base of the Tortonian has been defined at a level close to the Last Common Occurrence (LCO) of Globigerinoides subquadratus (Hilgen et al. 2000, 2005) with a new astronomical age of 11.625 Ma (Hüsing et al. 2007). It cannot be excluded that the Road section terminates into the lowermost Tortonian since Paragloborotalia siakensis at 4532 m (TR 120) is the only biostratigraphic marker species present above 4200 m.

In the Avcılar section, the occasional occurrences of Paragloborotalia siakensis up to 710 m (TR151) along with Globorotalia partimlabiata at 590 m and Globigerinoides subquadratus at 710 and 730 m, respectively, (TR 151 and 152) indicate that the lower 700 m of this section is also Serravallian in age. The absence of Globigerinoides
*subquadratus* and *Paragloborotalia siakensis* in the upper part of the Avcılar section along with the presence of *Globorotalia partimlabiata* near the top of the section (TR 169 and 172) suggests that the section extends up into the Tortonian.

The Avcılar section has been sampled assuming a significant overlap with the Road section and if the uppermost part of the Road section indeed extends into the Tortonian, we might assume an overlap of up to 1 km between these two sections. The maximum age range of the road section and Avcılar section is indicated by the age range of *Globorotalia partimlabiata* of 12.771–9.934 Ma (Hüsing et al. 2007).

**Paleoenvironmental interpretations for the Kahramanmaraş basin**

Deposition of the lower 200 m occurred in shallow marine conditions, as indicated by the occurrence of benthic foraminifers, calcareous nannofossils, poorly preserved echinoderms, gastropods and the large estuarine oyster *Crassostrea gryphoides* (Schlotheim, 1820). This large-sized Oligocene to Miocene species is restricted to brackish water environments with a high nutrient input and prefers building colonies on mud flats of outer estuaries (Slack-Smith 1998). Benthic foraminiferal species of the flysch succession from the road and Avcılar, such as *Cibicides* (*pseudo-*)ungerianus, *Gyroidina* spp., *Uvigerina* spp., *Oridorsalis* spp. and occasionally *Siphonina reticulata*, suggest water depths between 350 and 750 m during deposition of this section without evidence for shoaling towards the top, which is, in turn, cut by the thrust in the Avcılar section.

**Implications for the Kahramanmaraş basin**

During the Langhian – early Serravallian, shallow marine conditions prevailed in the Kahramanmaraş basin. The basin deepened during late Langhian/early Serravallian as indicated by the change from limestones and/or conglomerates to an alternation of marl and turbidites. Since neither in the lithology, nor in the biostratigraphic data, a shoaling trend towards the top of the section is observed, deep marine conditions (350–750 m) prevailed in the basin until the early Tortonian.

We interpret the whole section as a characteristic foreland basin flysch succession (as Dewey 1986; Karig & Közlu 1990; Derman & Atalık 1993; Derman 1999). Assuming the Road and Avcılar sections were sampled with no overlap, the maximum thickness is about 6.1 km, but assuming an overlap of up to 1 km, the maximum thickness is about 5.1 km. It is very difficult to estimate a sedimentation rate for this basin, because three sections were sampled with an unknown overlap. Secondly the accuracy of the age indicative biostratigraphic events is uncertain due to poor preservation and poor sampling resolution. Furthermore, the age indicative biostratigraphic events, LCO of *Globigerinoides subquadratus* and LO of *Globorotalia partimlabiata* are recorded in different sections, which makes the determination of the sedimentation rate between these two calibration points nearly impossible. The sedimentation rates thus vary much, between 50 and 450 cm/ka, but including slumps, debris flows and turbidites deposited in front of and during the activity of the thrust that now covers the top of the sequence. Taking a conservative estimate of 100 to 200 cm/ka, also because the LO of *Globigerinoides subquadratus* might not correspond to the true LCO, dated at 11.625 Ma (Hüsing et al. 2007), but might be higher in the stratigraphy, the age of the youngest flysch is about 11 Ma.

This age range, from Langhian to early Tortonian, differs significantly from the assigned Oligocene age of the open marine flysch and limestone deposits in the Muş and Elazığ basins. The continuous marine sedimentation in the Kahramanmaraş basin from Langhian to early Tortonian at a constant depth indicates that tectonic subsidence, possibly up to the order of 5 km, dominated the evolution of the basin.

**Discussion**

**Evolution of the east Anatolian basins**

The stratigraphic results from the east Anatolian basins are summarized in Figure 9, and are correlated to the Geological Time Scale (Gradstein et al. 2005). This figure schematically illustrates that the individual basins belong to two different, major basins: (1) a basin north of the Bitlis–Pötürge Massif, encompassing the Elazığ and Muş basins, which was filled with clastic mass flow deposits during the Rupelian and Chattian (Oligocene); (2) a basin south of the Bitlis–Pötürge Massif, a foreland basin which was filled with clastic sediments during the Langhian, Serravallian and early Tortonian (Middle and early Late Miocene).

The Muş and Elazığ basin, both north of the Bitlis–Pötürge Massif, show similar stratigraphic evolution during the Oligocene: Deepening of the basin occurred during the Rupelian and deep marine conditions (350–750 m) prevailed until the Chattian. Both basins evidence a shoaling trend during the Chattian. The macrofossil assemblage in the sandy limestones, such as molluscs and
echinoderm, indicates shallow marine, tropical to subtropical deposition, similar to a sheltered lagoon environment, with species preferring subtidal and intertidal environments. In addition, the macrofossil assemblage is comparable to assemblages found in Central Iran, the entire northern coast of the Western Tethys, the Central Paratethys and the lower Indus Ocean, indicating an open marine connection between these marine realms prior to the emergence during the late Chattian. We suggest
that the Muş and Elazığ basins were connected forming a large east–west elongated deep marine basin during Rupelian and Chattian.

The rapid deepening of the basin north of the Bitlis Massif may be related either to onset of flexural subsidence associated with (northward) underthrusting within the prevailing overall compressional regime (e.g. for the Elazığ basin Cronin (2000a, b)) or to the late stages of an extensional deformation period that persisted in the Paleocene and Eocene (e.g. in the Malatya basin (Kaymakçı et al. 2006) and in the Muş area (Şengör et al. 1985)). These two scenarios are controversial and our data provide age and depth constraints on the Muş and Elazığ basins, which do not allow to eliminate or prefer either of these scenarios.

In the Kahramanmaraş basin, south of the Bitlis–Pötürge Massif, shallow marine sediments were deposited during the Langhian. A rapid deepening during the Langhian to Serravallian indicated by the rapid transition to deep marine (350 to 750 m) flysch deposits, was followed by deposition of continuously deep marine sediments until the early Tortonian. Since no shoaling trend is observed we suggest that the age of the youngest flysch underneath the thrust, biostratigraphically dated as early Tortonian, at about 11 Ma, coincides with the end of underthrusting.

The rapid deepening of the foreland basin south of the Bitlis–Pötürge Massif during the Langhian to Serravallian, followed by the deposition of a thick deep marine flysch succession, can be interpreted as northern Arabia and more specifically the area of the Kahramanmaraş basin, entering into the subduction zone underneath Anatolia. The end of flysch deposition and thus the youngest flysch underneath the thrust of the overriding Bitlis–Pötürge Massif could be envisaged as the end of subduction, thus underthrusting, at about 11 Ma (Tortonian), which is likely followed by rapid uplift in the region. Such episodes of very rapid uplift and folding of foreland basins associated with the stalling of underthrusting is, for example, also well documented in the western Aegean region (Richter et al. 1978; van Hinsbergen et al. 2005a, c, d).

Our new results of the Kahramanmaraş basin can be compared to previously published data from the Hatay (around Antakya) and Lice regions (see Figs 2 & 9), which have been interpreted as foreland basins related to southward thrusting of the Taurus allochton over the Arabian continental margin belonging to an east–west elongated foreland basin overlying the Arabian promontory (e.g Perińçek 1979; Karig & Kızılca 1990; Derman and Atalik 1993; Derman 1999; Robertson et al. 2004).

The stratigraphy and chronology of the Hatay area is very similar to the evolution of the Kahramanmaraş basin. The chronology in the Hatay area has recently been redefined based on micropaleontological dating (Boulton et al. 2007) and we can therefore correlate the evolution of the Kahramanmaraş basin to the Hatay area. The stratigraphy in the Hatay area is characterized by a pronounced angular unconformity between middle Eocene and overlying lower Miocene sediments, with a hiatus in the Oligocene (Boulton & Robertson 2007). Sedimentation resumed during the Aquitanian to Burdigalian (Early Miocene) with deposition of conglomerates and mudstones. In the Kahramanmaraş area, Derman & Atalik (1993) and Derman (1999) assigned a lower Miocene age to the about 1 km thick series of fluvial deposits, which precede the thick flysch deposits. We, however, have no age constraints on the fluvial deposits and can therefore not confirm an Early Miocene age. During the Langhian both basins experienced shallow marine limestone deposition and the basin progressively deepened during Serravallian to Tortonian (Boulton et al. 2007; Boulton & Robertson 2007). The deposition of shallow marine limestones in the Hatay area have been interpreted to be related to further loading of the lithosphere in response to flexural subsidence and the progressive deepening to flexural control (Boulton & Robertson 2007). Where flexural subsidence exceeded the build up of a carbonate platform hemipelagic sediments were deposited (Boulton et al. 2007; Boulton & Robertson 2007). A similar scenario can be envisaged for the Kahramanmaraş basin indicated by coarsening upwards in the thick flysch deposition. In the Hatay area, by the end of the Miocene, the tectonic regime changed and the Pliocene–Quaternary Hatay Graben structure was formed in a transtensional setting related to the EAF (Perińçek & Cemen 1990; Boulton et al. 2007; Boulton & Robertson 2007), while deep marine sediments in the Kahramanmaraş basin were overthrust already during the early Tortonian. This comparison might indicate a diachronous evolution of these two basins with the Kahramanmaraş basin emerging during the Tortonian and the Hatay area remaining open marine until the deposition of Messinian evaporites, or different basin evolutions due to the relatively western position of the Hatay area thus closer to the present-day extent of the eastern Mediterranean.

A comparison to the Lice basin, which is situated to the east of the Kahramanmaraş basin, would evidently give constraints on the syn- or diachronous evolution of the southernmost, Arabian foreland basin. However, the chronology of sediments in the Lice basin is scarcely documented in the literature (e.g. Perińçek 1979; Dewey 1986; Karig & Kızılca 1990; Robertson et al. 2004). On the geological map of Turkey (Şenel 2002) shallow marine clastic and carbonatic sediments have been indicated as Early Miocene in age and continental...
clastic rocks as Middle to Late Miocene in age. This succession would pre-date the flysch deposition in the Kahramanmaras and Hatay area and would indicate diachronous evolution of the elongated Arabian foreland basin. Other studies assigned, however without documenting an age control, a Tortonian age to the Lice flysch (Dewey 1986). If the flysch deposits in the Lice, Kahramanmaras and Hatay area are indeed synchronous, we would assume a synchronous evolution of the Arabian foreland basin which emerged during the Tortonian. However since the chronology of the Lice basin is not well documented, firm correlation to the Kahramanmaras basin and Hatay area remains impossible (see question marks in Fig. 9).

The basin south of the Bitlis–Pötürgê Massif including the Kahramanmaras, Hatay and Lice basins, is interpreted as the southernmost and youngest foreland basin in the east Anatolian fold-and thrustbelt, which formed as a large east-west trending foreland basin on the subducting Arabian plate. The end of underthrusting in the Kahraman maras basin is dated at about 11 Ma, but might have been diachronous relative to the emergence of the Hatay and Lice basin.

**Tectonic closure of the eastern Tethys gateway**

Based on the presented data herein, we envisage the following scenario for the Oligocene to Miocene evolution of the basins north and south of the Bitlis Massif in SE Turkey (Fig. 10). During the early Oligocene, marine sediments were deposited in a large basin to the north of the Bitlis–Pötürgê Massif (Fig. 10a). However, our data does not allow us to constrain whether the deepening of

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**Fig. 10.** The evolution of the Oligocene–Miocene basins in SE Turkey are illustrated schematically in three major phases: (a) during the Oligocene from c. 30 to c. 23 Ma: a marine basin was situated north of the Bitlis–Pötürgê (BP) until the end of latest Oligocene, Chattian, when this basin emerged; (a1) related to extension, (a2) related to thrusting. (b) during the Langhian to early Tortonian (c. 13 to c. 11 Ma): areas of present-day northern Arabia enter the position of the foreland basin South of the BP and the northern Arabian promontory was subducted underneath the BP from the Langhian until the early Tortonian, c. 11 Ma, and finally. (c) since the early Tortonian to Recent: the end of large-scale underthrusting at c. 11 Ma in east Anatolia coincides with the onset of collision-related volcanism, uplift of the East Anatolian Accretionary Complex (EAAC), the onset of shearing along the North and East Anatolian Faults (NAF and EAF). Refer to text for further discussion. BP = Bitlis–Pötürgê; BPM = Bitlis–Pötürgê Massif; NAF = North Anatolian Fault; EAF = East Anatolian Fault; EAAC = East Anatolian Accretionary Complex; KM = Kahraman maras.
the basin during the Rupelian was related to either large scale extension (Fig. 10a1) or thrusting (Fig. 10a2), with the Bitlis–Pötuşge Massif situated on the overriding plate. Nevertheless, our data suggest that until the Chattian, flysch was deposited in a deep marine environment, as recorded in the area of Muş and Elazığ. The emergence of the basin north of the Bitlis–Pötuşge Massif during the late Chattian (see also Fig. 9) probably coincides with the accretion of the Bitlis–Pötuşge Massif to the Anatolian plate.

On the southern side of the Bitlis–Pötuşge Massif, oceanic subduction was probably ongoing due to Africa/Arabian’s relative distal position (e.g. Besse & Courtillot 2002).

During the Langhian to Serravallian the basin south of the Bitlis–Pötuşge Massif deepened rapidly, which might be related to the northern Arabian promontory, the present-day northern margin of the Arabian plate (Kahramanmaraş, Hatay and Lice basins), entering into the subduction zone (Fig. 10b) below the Bitlis–Pötuşge Massif. During the Serravallian and early Tortonian, the Kahramanmaraş basins remained deep marine indicated by thick flysch deposition until, at least, the early Tortonian. The youngest flysch underneath the thrust in the Kahramanmaraş area, biostratigraphically dated at about 11 Ma, might be linked to the end of the large-scale underthrusting (subduction) in eastern Anatolia (Fig. 10c). In models proposed by Keskin (2003) and Şengör (2003), it is assumed that the Bitlis–Pötuşge Massif was accreted with Arabia during late Eocene, while Robertson et al. (2004) suggested Late Oligocene–earliest Miocene time. Our data, on the other hand, indicate the presence of a deep marine realm between the Bitlis–Pötuşge Massif and Arabia during Serravallian and early Tortonian, which we suggest is associated with the continuous subduction of Arabia underneath the Anatolian plate.

The timing of the end of thrusting agrees with the onset of the collision-related volcanism at about 11 Ma north of the present-day suture line (Keskin 2003), the uplift of the East Anatolian Accretionary Complex inferred to start around 11 Ma onwards (Şengör et al. 2003) and the onset of the North and East Anatolian Fault (Dewey 1986; Hubert-Ferrari et al. 2002; Şengör et al. 2005) (Fig. 10c). Collision-related volcanism and uplift of the East Anatolian High Plateau despite the relatively thin crust (45 km) has been related to an anomalously hot mantle underneath the eastern Anatolia (Keskin 2003; Şengör et al. 2003). This, as well as the onset of westward extrusion of Anatolia and the onset of formation of the North and East Anatolian faults, have been explained by slab detachment at about 11 Ma in eastern Anatolia (Keskin 2003; Şengör et al. 2003; Şengör et al. 2005; Faccenna et al. 2006), which is in line with a recent tomography study of Hafkenscheid et al. (2006). Our new results from the Kahramanmaraş area can thus be considered in line with the previously suggested scenarios, the end of underthrusting and the onset of extrusion of Anatolia in the Late Miocene (at about 11 Ma) (Fig. 10c).

Constraints on the closure of the eastern Tethys gateway

The continuous northward migration of the African–Arabian plate led to the disruption of the Tethys seaway and the final closure related to continental collision of Arabia and Eurasia. The paleogeographic extent of the Tethys during the Paleogene and Neogene thus underwent significant changes until the connection was finally closed. Several authors suggested that the final closure of the eastern Tethys gateway may have resulted in significant changes in the paleoceanographic circulation and consequently in a major change in global climate (e.g. Woodruff & Savin 1989, 1991; Jacobs et al. 1996; Flower & Kennett 1993; Yılmaz 1993).

Our data from eastern Anatolia indicate that a deep marine connection was present north of the Bitlis–Pötuşge Massif from Rupelian to late Chattian. The shoaling of this northern basin during the late Chattian led to severe disruption between an eastern (Indian Ocean) and western (Mediterranean) marine domain; particularly the deep-water circulation was disrupted during the Chattian. The emergence of this basin after the late Chattian resulted in the closure of at least this branch of the southern Tethys and coincides with the late Oligocene warming, reducing the extent of the Antarctic ice, which was punctuated by the Mi-1 glaciations around the Oligocene–Miocene boundary (Zachos et al. 2001).

Other studies suggest that the Tethys seaway was open until the early Miocene and became severely restricted during the Burdigalian (c. 19 Ma), when mammal fauna and shallow marine macrofaunal records from the eastern Mediterranean region indicate the existence a landbridge (Gomphotherium landbridge) connecting Africa/Arabia and Eurasia (Popov 1993; Rögl 1998, 1999; Harzhauser et al. 2002, 2007). These authors claim that, since c. 19 Ma, biogeographic separation between the Mediterranean-Atlantic and Indo-Pacific regions persisted; despite some short-lived periodic marine connections between the two domains until the middle Miocene (Rögl 1999; Meulenkamp & Sissingh 2003; Golonka 2004; Harzhauser et al. 2007). If a causal link between the closure of the eastern Tethys gateway and global climate cooling exists, a major change in global, or at least local, climate must be expected during the Burdigalian time. The most significant climatic change during the Burdigalian, as evidenced in both the δ18O and
\[ \delta^{13}C \] record, indicates a change from a cooling to a warming trend which led into the Mid-Miocene Climatic Optimum (Zachos et al. 2001).

Our data suggest that if a deep marine connection between the eastern and western marine realm persisted after the late Chattian, it was probably located south of the Bitlis–Pötürge Massif. The studied basins along the south Bitlis suture zone in eastern Anatolia, however, do not comprise the stratigraphic interval between late Chattian and Langhian (25–15 Ma). Consequently, it is not possible to constrain the tectonic evolution and the palaeogeographic extent of the Tethys seaway during this time interval from the stratigraphic record of the east Anatolian basins.

In the context of global climate change, the main oxygen and carbon isotope shift corresponds to the second and major step (Mi2b) of the middle Miocene global cooling, and has recently been astronomically dated at 13.82 Ma, close to the Langhian–Serravallian boundary, in a section on Malta (Abels et al. 2005). The middle Miocene decrease in \( \delta^{18}O \) values was previously attributed, amongst other hypotheses, to a possible local expression of the isolation of the Mediterranean Sea from the Indo-Pacific Ocean (van der Zwaan & Gudjonsson 1986; Jacobs et al. 1996). Abels et al. (2005), however, show that this event coincides with a period of minimum amplitudes obliquity related to the 1.2-Ma cycle and minimum values of eccentricity as part of both the 400– and 100-ka eccentricity cycle, thus suggesting astronomical forcing (see Abels et al. 2005).

If a link between gateway closure and middle Miocene climate change exists, the south Bitlis gateway must have re-opened to finally close in the middle Miocene, which is very unlikely in an overall converging setting. Moreover, our data does not show evidences for a final closure of the seaway in the middle Miocene. In contrast, the data from Kahramanmaraş indicates rapid deepening during the Langhian to Serravallian and prevailing deep marine conditions until the early Tortonian. This has been interpreted as related to continuous northward subduction underneath the Bitlis–Pötürge Massif and finally continental collision during the Serravallian to early Tortonian. Our data suggest that a deep marine connection located between the Bitlis–Pötürge Massif and Arabia, whether periodic or not, was disrupted at latest during the early Tortonian, giving an upper limit of c. 11 Ma to the final closure between the Indian Ocean and the Mediterranean along the northern Arabian. The above analysis shows that the end of foreland basin existence in SE Turkey – and therefore the closure of the southern Tethyan gateway – can not straightforwardly be linked to the middle Miocene climate change. Future assessment of the timing of the Tethys gateway closure should focus on detailed stratigraphy of the youngest foreland basins in SE Turkey, NW Iran, Syria and N Iraq, the region of the Bitlis–Zagros suture zone.

Conclusions

The marine basin north of the Bitlis–Pötürge Massif encompassing the Elazığ and Muş basins emerged during Chattian, which was followed by shallow marine limestone deposition during the late Chattian and finally closed after the late Chattian. This marks the disruption of the Tethys gateway north of the Bitlis–Pötürge Massif connecting an eastern (Indian Ocean) and western (Mediterranean Sea) domain.

The Kahramanmaraş basin south of the Bitlis–Pötürge Massif, probably linked to the Hayat and Lice basins, experienced shallow marine conditions during Langhian, rapidly deepened during Langhian to Serravallian and remained deep marine during the Serravallian and early Tortonian. No shoaling trend has been observed in the Kahramanmaraş basin and the age of the youngest flysch underneath the subduction-related thrust has been biostratigraphically dated at early Tortonian, at about 11 Ma. The end of flysch deposition in the Kahramanmaraş area is probably related to the end of subduction, thus the end of underthrusting. The age coincides with the onset of collision-related volcanism, uplift of the East Anatolian Accretionary Complex, and the timing of shearing along the NAF and EAF. Our new results suggest a strong link between the processes outlined above, which have been explained by slab detachment at about 11 Ma in eastern Anatolia.

This age, early Tortonian, about 11 Ma, is the youngest possible age for a deep marine connection between the Mediterranean-Atlantic and Indo-Pacific regions. We can thus constrain the timing of the final closure of a deep marine Tethys gateway to an upper limit of about 11 Ma. The emergence of the basin north of the Bitlis–Pötürge Massif during the late Chattian thus provides a lower limit of the closure of the eastern Tethys gateway.

In the southern basins marine foreland deposition was continuous during Serravallian and early Tortonian and our data does not support a link between the Middle Miocene climate cooling dated at 13.82 Ma and the closure of the eastern Tethys gateway. In contrast, the age of the youngest flysch deposits, thus the youngest foreland basin in SE Turkey is early Tortonian, about 11 Ma.

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