Int J Earth Sci (Geol Rundsch) (2016) 105:339–368
DOI 10.1007/s00531-015-1248-1

ORIGINAL PAPER

Miocene transgression in the central and eastern parts of the Sivas Basin (Central Anatolia, Turkey) and the Cenozoic palaeogeographical evolution

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Received: 21 January 2015 / Accepted: 13 September 2015 / Published online: 16 November 2015

Abstract We present here a reappraisal of the tectonic setting, stratigraphy and palaeogeography of the central part of the Sivas Basin from Palaeocene to late Miocene. The Sivas Basin is located in the collision zone between the Pontides (southern Eurasia) and Anatolia (a continental block rifted from Gondwana). The basin overlies ophiolites that were obducted onto Anatolia from Tethys to the north. The Central Anatolian Crystalline Complex (CACC) experienced similar ophiolite obduction during Campanian time, followed by exhumation and thrusting onto previously emplaced units during Maastrichtian time. To the east, crustal extension related to exhumation of the CACC created grabens during the early Tertiary, including the Sivas Basin. The Sivas Basin underwent several tectonic events during Paleogene–Neogene. The basin fill varies, with several sub-basins, each being characterised by a distinctive sequence, especially during Oligocene and Miocene. Evaporite deposition in the central part of the basin during early Oligocene was followed by mid-late Oligocene fluvio-lacustrine deposition. The weight of overlying fluvial sediments triggered salt tectonics and salt diapir formation. Lacustrine layers that are interbedded within the fluviatile sediments have locally yielded charophytes of late Oligocene age. Emergent areas including the pre-existing Sivas Basin and neighbouring areas were then flooded from the

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east by a shallow sea, giving rise to a range of open-marine sub-basins, coralgal reef barriers and subsiding, restricted-marine sub-basins. Utilising new data from foraminifera, molluscs, corals and nannoplankton, the age of the marine transgression is reassessed as Aquitanian. Specifically, age-diagnostic nannoplankton assemblages of classical type occur at the base of the transgressive sequence. However, classical stratigraphic markers have not been found within the planktic foraminiferal assemblages, even in the open-marine settings. In the restricted-marine sediments, there are rich planktic foraminiferal assemblages of classical type but these are of little use in stratigraphy. In contrast, the gastropod fauna indicate a Burdigalian age. Sediment reworking in the restricted-marine environments precludes stratigraphic determination. In such environments, micro- and nano-organisms experienced atypical developmental conditions. The small benthic foraminifera and associated ostracod assemblages are good indicators of salinity which varied considerably within the restricted-marine sub-basins. Some of the corals within the coralgal reefs barriers are also dated as Aquitanian. A combination of the salt tectonics and the late Miocene north-westward-verging thrusting created the present basin complexity.

Keywords Miocene · Stratigraphy · Cenozoic palaeogeography · Sivas Basin · Turkey

Introduction

After the closure of the Northern branch of Neotethys in late Cretaceous times, several intracontinental basins developed around the Central Anatolian Crystalline Complex (CACC, Göncüoğlu et al. 1991). These are, from the north-west to the south-west of the CACC, the Çankırı, Haymana, Tuz Gölü and Ulukışla (to the south of Nigde) Basins. The Sivas Basin is located to the SE of the CACC and to the north of the Taurus (Fig. 1). These basins began as marine, more or less deep, subsident basins in a regional context of N–S shortening and the convergence of the Arabian promontory, the Taurus and Kırşehir blocks and the Eurasiatic plate.

The Sivas Basin, which is the largest of these basins (250 km from west to east and about 50 km from north to south), extends from the region of Kayseri westwards, to Erzincan in the east. The basin covers the Kırşehir massif along its north-western border (to the west of Sivas) and the Northern Neotethyan suture (to the east of Sivas). To the south, it covers the Taurus belt. The Sivas Basin rests on the ophiolites of the Ankara–Erzincan suture, which were obducted southwards during late Cretaceous time over the Kırşehir Massif and also over the eastern margin of the Taurus belt. The basin experienced several north–south extensional and compressional phases and, as a result, records the collisional and post-collisional history of the Alpine belt in Central and Eastern Anatolia; it is thus important to help understand the geodynamic evolution of the Middle East during Cenozoic time. With regard to the closure of the Northern branch of Neotethys in late Cretaceous times, the Sivas Basin can be considered as a typical post-tectonic basin (Yılmaz 1994). However, it is also a syn-tectonic basin which has recorded a succession of tectonic events (compressional end extensional) which affected its continental basement during Tertiary times. Each of these events played an important role in the variation of the geometry and infill of the basin as discussed below.

The present structural setting of the basin is essentially the result of the final compressional event which generated a prominent system of thrusts, known as the "Sivas back thrust" along which the basin has been thrust towards the N–NW onto the late Miocene-Pliocene (?) Incesu Formation. This deformation affected the northern margin of the basin, near Sivas, the margin of the Kırşehir massif in the west (it has been recognised as far as Kayseri to the west) and the Northern Neotethyan suture to the east of Sivas. This compressional event is, therefore, a major tectonic event in the Sivas Basin (Poisson et al. 1992, 1996) (Fig. 2). As a result, the main sedimentary sequence of the basin was structurally repeated several times, with gypsum beds acting as a décollement. In addition, pre-existing structures could have been hidden or reoriented by thrusting. This is particularly the case for diapiric structures and also for the Deliler Fault which separates the basin longitudinally into a southern sub-basin and a northern sub-basin (Cater et al. 1991).

These fundamental structural features were previously known, but their role in basin evolution was not well explained. This is particularly the true for the diapiric structures and the related salt tectonics which are now being restudied (Ringenbach et al. 2013; Callot et al. 2014; Ribes et al. 2015).

Below, we discuss the Cenozoic palaeogeographical evolution of the Sivas Basin and its neighbouring areas on the basis of new biostratigraphic data concerning foraminifera, nannoplankton and mollusc groups, and also supplementary K/Ar dating.

Previous work

Detailed geological study of the Sivas Basin largely began after the creation of the Maden Tetkik ve Arama Enstitüsü (MTA) in 1935. Initial studies mainly concerned lithostratigraphic and biostratigraphic description. Chaput (1936), while travelling Sivas to Diyarbakır, dated the upper part of the Işhanı bioclastic limestones near Sivas as late Burdigalian in age. Further south, he attributed lacustrine limestones of the southern border of the basin to the late

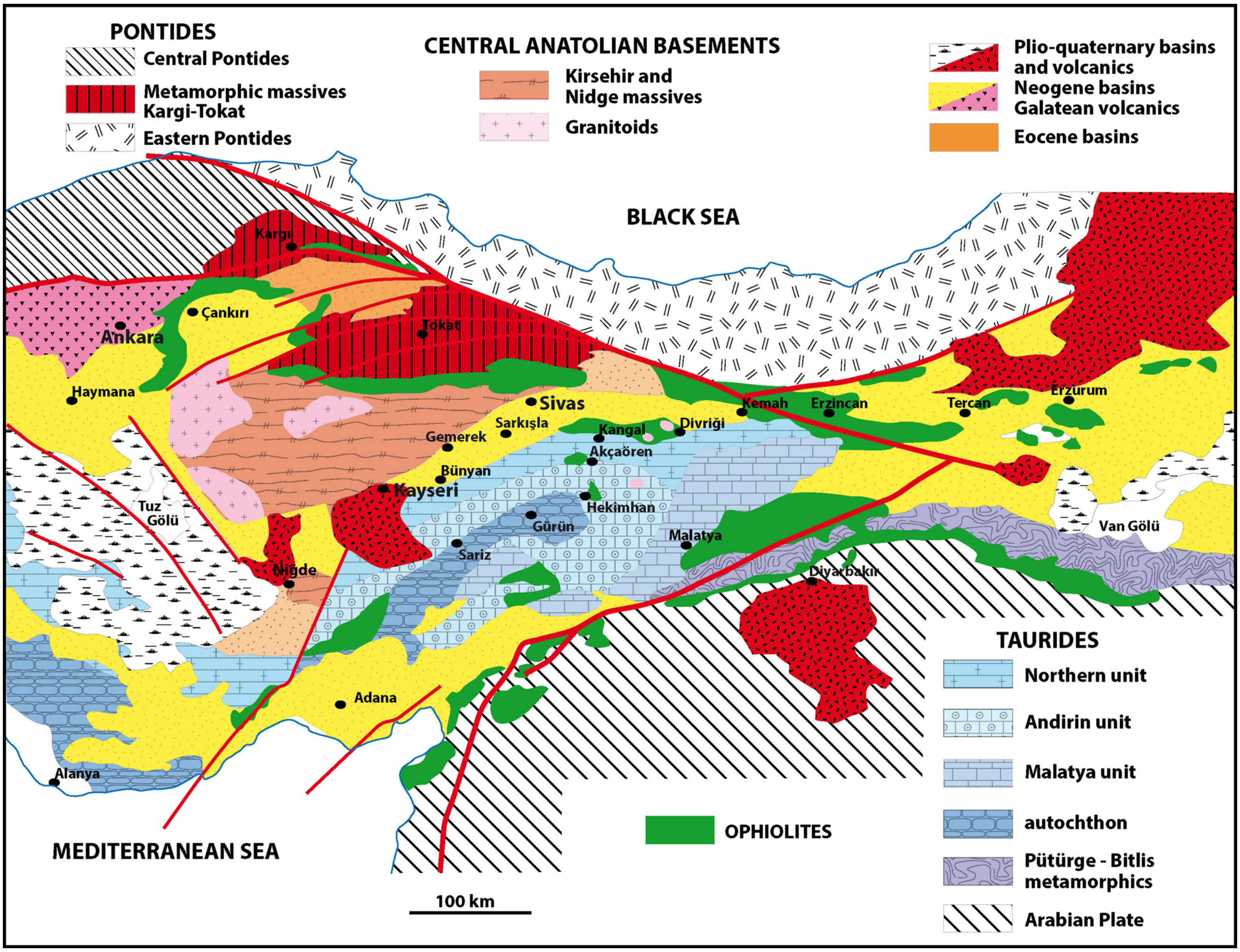

Fig. 1 Location of the Sivas Basin in the Middle East area

Oligocene. Stchépinsky (1939) later carried out a detailed study of the marine fauna, mostly molluscs, and some echinids and corals. He recognised the transgressive character of shallow-marine Lower Miocene sediments above thick Oligocene evaporitic and red clastic sequences. He also provided the first geological map of this area.

Geological studies recommenced after the Second World War. The first general synthesis of the Sivas Basin was coordinated by Baykal and Erentöz (1966), leading to the publication of the 1/500,000-scale Sivas sheet. This remained the reference map for the basin until publication of the much more recent 1/100,000 scale maps for the area around Sivas (Aktimur 1988a, b; Aktimur et al. 1990; Yılmaz et al. 1998).

Concerning structural geology, Baykal and Erentöz (1966) pointed out the existence of NE–SW trending folds and of an E–W southward-verging thrust at the northern margin of the basin, supposedly of post-Oligocene to pre-Miocene age. We have not observed such a thrust. However, perhaps this structure could relate to the south-verging Central Anatolian Thrust Belt which resulted from the closure of the Northern Neotethys during late Cretaceous–Palaeocene times (Figs. 1, 2). Northward-directed reverse faults and thrusts have been reported from the southern margin of the basin (Tecer Dağ) by Arpat (1964), Artan and Sestini (1971) and Kurtman (1973). Kurtman (1973) interpreted the Sivas thrust as a left-lateral strike-slip fault. For Aktimur (1988a) this was a south-verging, left-lateral transpressional strike-slip fault, which, according to Cater et al. (1991), experienced earlier down-to-the-south extensional displacement. The Sivas thrust has been also considered as a strike-slip fault related to the Central Anatolian Fault Zone (CAFZ), which is a prolongation of the Ecemiş Fault (Koçyiğit and Beyhan 1998). We have not observed any evidence of major strike-slip movement along the Sivas thrust fault zone, which is well constrained kinematically as indicating compression towards the N–NW (Poisson et al. 1992, 1996). In addition, the Deliler fault zone in the area (Fig. 12a Palaeocene–early

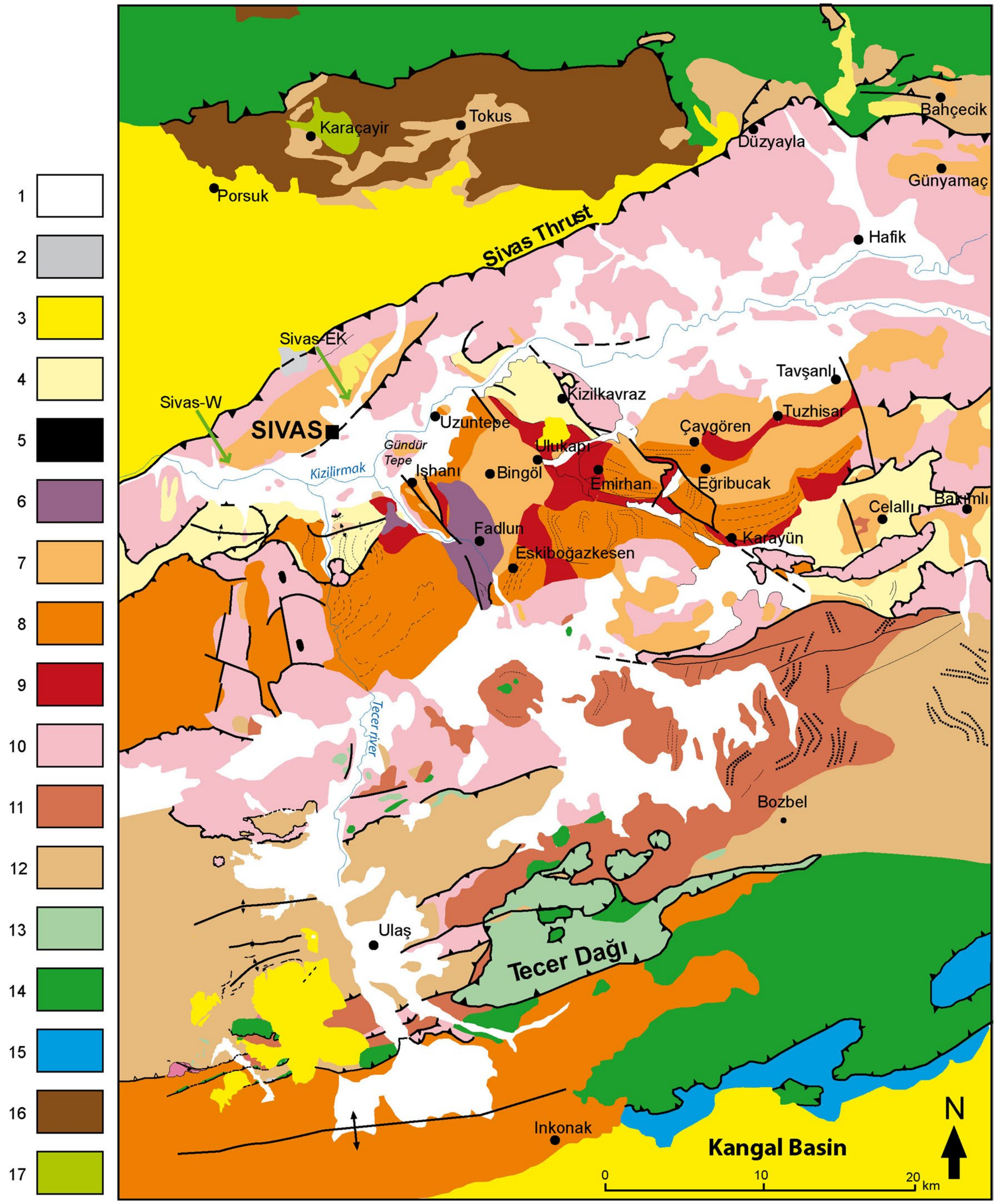

Fig. 2 Geological map of the central part of the Sivas Basin (modified after Poisson et al. 1996). *1* Quaternary; *2* travertine; *3* late Miocene-Pliocene (?) lacustrine limestones (Merakom Formation); *4* Benlikaya Formation (early to middle (?) Miocene), around Celallı and Kızılkavraz; Incesu Formation (late Miocene), to the south of Sivas; *5* mid-Miocene basalts; *6* Fadlun (early to middle (?) Miocene); *7* early Miocene marine to lagoonal sequences of the Bingöl, Eğribucak, Karayün, Celallı and Ağılkaya minibasins; *8* mid- to late Oligocene red clastic sequences of the minibasins (Karayün Formation); *9* gypsum diapirs in the diapir area east of Sivas; *10* early Oligocene gypsum (Hafik Formation); *11* Oligocene red clastic sequence (Selimiye Formation); *12* Eocene flysch (Bozbel Formation); *13* Maastrichtian–Palaeocene shallow-marine limestones of the Tecer Dağ; *14* late Cretaceous Northern Neotethyan ophiolitic nappes and ophiolitic mélange; *15* Mesozoic Taurus carbonate Platform; *16* Kırşehir Massif; *17* Karaçayır intrusive syenite (100 Ma). Sivas-W and Sivas-EK: location of the sections referred to in the text

Eocene time interval) corresponds to a system of deep faults along which submarine volcanism was active during Palaeocene and Eocene times (Gökten 1986; Temiz 1996). Cater et al. (1991) related this fault to the initiation of two adjacent longitudinal sub-basins which appeared to have experienced different geodynamic evolutions.

Concerning regional geology, the southern part of the Sivas Basin was considered in a number of publications, specifically the Tecer Dağ and the area between the Tecer Dağ and Divriği (Arpat 1964; Artan and Sestini 1971; Gökten 1978, 1993; Inan 1987, 1988; Inan and Inan 1987; Inan et al. 1993; Kangal and Varol 1999). In addition, Miocene sediments were studied by Kavak et al. (1997) and by Kavak (1998). The stratigraphy of the northern margin of the basin (around Hafik and Zara) was also revised (Gökçen 1981, Gökten 1978, 1993; Gökten and Kelling 1991). For the Celallı area (south of Hafik), Gökçen and Kelling (1985) gave a sedimentological interpretation of the Selimiye red clastics and dated them as Oligocene for the first time on the basis of an ostracod microfauna. For a long time this remained the only dating of the Selimiye Formation. Despite these localised advances, the l970s synthesis of Kurtman (1973) remained the most complete description of the litho-stratigraphic formations and their faunal content in the eastern half of the Sivas Basin, the main area discussed below. Kurtman (1973) also mapped the main tectonic lineaments in the area, such as the NE–SW trending folds and some thrust faults, and also recognised the existence of diapirs. In addition, the western part of the Sivas Basin was studied by Gökten (1983, 1985, 1986), who mapped the area around Şarkışla, described the sequences and also dated them. In this area the general sequence was shown to be composed of Palaeocene–Eocene deep-marine deposits including important basaltic lava flows and volcaniclastic breccias. This volcanism was interpreted as the result of an extensional phase that gave rise to normal faults within the central part of the basin. Andesitic and basaltic lavas poured out onto the sea-floor along fault zones as pillow basalts that were in places reworked to form submarine sedimentary breccias (Gökten and Floyd 1987). A Palaeocene phase of extension was thus documented for the first time in the Sivas Basin. We can now show that this extensional phase remained active until the late Eocene, as seen in the Ortaköy area (Fig. 4). An olistostrome containing ophiolitic rocks and sedimentary blocks was also described by Gökten and Floyd (1987). Ophiolitic rocks, basaltic lavas and large blocks of shallow-marine limestones of Maastrichtian–Palaeocene–early Eocene age (Tecer Dağ limestones) appear to have been exhumed from the basement along the Deliler fault. Further studies of the south-central part of the basin (Gökten 1993) extended the data of Kurtman (1973) and showed that sedimentation during the Eocene was relatively uniform throughout the Sivas Basin, being made up of deep-marine flysch, intercalated volcanic products and olistostromes. In addition, continental formations which occupy large areas and thicknesses of the Sivas Basin were dated at a small number of sites utilising of vertebrate fossil assemblages and plant remains (pollens and charophytes). These fossils occur in lacustrine deposits, marls and limestone lenses, interstratified with the fluviatile sediments (red pelites, sandstones and conglomerates), and also in lacustrine sediments above "massive gypsum". The vertebrates were dated as middle and late Oligocene, Miocene and Pliocene in the western part of the basin (Sümengen et al. 1990), as late Miocene and Pliocene in the Incesu Formation to the NW of Sivas (Yalçınlar 1955) and to the north of Hafik, and also as late Oligocene along the southern border of the basin near Inkonak (De Bruijn et al. 1992; Ünay et al. 2003). Charophyte assemblages have been attributed to the Oligocene and Miocene (Poisson et al. 2012). Pollen analysis has been performed for some localities giving a middle Miocene age (Atalay 1993, 1999). For other areas biostratigraphic data were revised and completed in stages, for example, for molluscs (Erünal-Erentöz 1956; Poisson et al. 1997), planktic foraminifera (Poisson et al. 1997; Özden et al. 1998), benthic foraminifera (Dizer 1962; Poisson et al. 1997; Suata and Inan 1998; Özden et al. 1998; Özcan et al. 2009) and nannoplankton (Çubuk and Inan 1998). A recent study concerned Oligocene benthic foraminifera, including the description of new species from the central part of the basin (Sirel et al. 2013).

Several studies concerned the palaeogeography and the palaeoenvironments of the Sivas Basin. Cater et al. (1991) proposed a subdivision into southern and northern sub-basins. These authors discussed the syntectonic evolution of the basin and the role of the evaporites. However, they did not revise the stratigraphy. They compared the Hafik evaporites to the Mediterranean Messinian evaporites which was incorrect. The sequences in two areas of the basin: to the east, near Imranli and in the central part of the basin to the SE of Sivas (Karayün section), were described by Çubuk (1994) and by Çubuk and Inan (1998). These authors recognised the Emirhan diapiric structure. Cyclic sedimentation and palaeoenvironments in the central part of the Sivas Basin during the Oligo-Miocene were described in detail for the first time by Çiner et al. (1995, 2002) and by Kosun (2002). These authors established a southerly origin for the clastic material in the basin and mapped gypsum diapirs. In their model, the Ağılkaya and Eğribucak Formations represent a continuous stratigraphic succession. Recent data, however, show that the base of the Eğribucak Formation (Oligocene) is older than the top of the Ağılkaya (Karayün) Formation (Miocene) (Sirel et al. 2013). Rather than being stratigraphically one above the other, these two formations are instead interpreted as lateral equivalents that are separated by a previously inferred thrust fault (Poisson et al. 1996; Ribes et al. 2015).

Ongoing work on the Sivas Basin concerns the sedimentology and the tectonics, especially the salt tectonics (Callot et al. 2014), which was previously neglected. The existence of some diapirs had been recognised (Chaput 1936; Kurtman 1973; Cater et al. 1991), mapped and briefly described (Çubuk and Inan 1998; Çiner et al. 2002). However, the existence of minibasins between the diapirs, their halokinetic sequences and the associated tectonic structures were not taken into account (Ribes et al. 2015; Kergaravat, in preparation). These studies considerably improve models of the sedimentation and tectonic evolution above and around the diapirs. The studies also help to constrain tectonic, palaeogeographic and palaeoenvironmental reconstructions at the scale of the Sivas Basin as a whole.

In addition, geophysical study of the Sivas Basin is a recent development. An electric profile has been published (Önal et al. 2008), which, however, remains difficult to interpret. More recently, a seismic survey has been carried out in the central part of the basin, but the results have not been published.

Litho-stratigraphic units and revised stratigraphy

A single log cannot be given for the Sivas Basin as a whole because of local sedimentary variation (see Fig. 3). The initial formal descriptions of the formations by Kurtman (1973) were later developed by several authors who described local sections in more detail. A synthesis, with a reappraisal of the biostratigraphy, was given by Poisson et al. (1996, 1997). Kurtman (1973) described several of the main formations in the eastern half of the basin which he considered to be valid for the entire Sivas Basin. This is appropriate for only some of the formations, however. One of the most important formations, the Karacaören Formation has since been subdivided into smaller units based on regionally important changes in facies and new stratigraphic data. For the western half of the Sivas Basin, the general succession, from bottom to top, can be revised as follows:

Bahçecik Conglomerates (Kurtman 1973) (synonym: Özderesi Formation; Gökten and Kelling 1991)

This formation has a relatively local distribution at the northern margin of the Sivas Basin, near the village of Bahçecik (north of Hafik), where it rests directly on the Central (or North) Anatolian Thrust Belt. The formation corresponds to the first marine deposits on the emergent Northern Neotethyan nappes along the northern border of the Sivas Basin. The sequence consists of coarse-grained, thick-bedded polymict conglomerates that accumulated in a fluvio-marine fan-delta setting. Cross-bedding and channel scours are common. Kurtman (1973) inferred an early Eocene age for the formation. Later the base of the formation was dated by Gökten and Kelling (1991) and by Poisson et al. (1996) as late Palaeocene–early Eocene, utilising gastropods (*Batillaria* sp.). The uppermost part of the formation consists of nummulitic sandstones and marls of early Lutetian age, which can be correlated with the Bozbel Formation (see below). According to Kurtman (1973), the Bahçecik Conglomerates reach 1500 m thick, although estimated as only 300 m thick by Gökten and Kelling (1991). Further west, similar deposits, known as the Tokuş Formation, occur in the Karaçayır area, in the Tokuş area (Özden et al. 1998), and also on the SE margin of the Kırşehir Massif (to the north of Gemerek).

Bozbel Flysch (Kurtman 1973)

This facies crops out mainly along the southern margin of the basin (the southern basin of Cater et al. 1991), where it has been tectonically exhumed. The type locality is in the south of the basin although the formation has a large extension from west (Ortaköy area; Fig. 3) to east (Kemah-Erzincan area). Sandstones and marls of turbiditic origin are intercalated with volcanogenic layers (lava and tuffite of Palaeocene and Eocene age; Fig. 4); there is also an olistostrome made up of limestone blocks of Lutetian age (Artan and Sestini 1971; Gökten 1983, 1985, 1986, 1993). The sandstones are composed of quartz, feldspar, volcanic rocks, serpentinite, radiolarites and resedimented calcarenites with nummulites. The age of the formation is Palaeocene to late Eocene, and the terrigenous components are believed to have been derived from the south. The thickness is at least c. 2000 m.

Hafik Formation (Kurtman 1973)

The formation is composed of a sequence of layered gypsum in the form of thick, large lenses which crop out all along the northern margin of the Sivas Basin, from Gemerek to Imranlı and eastwards, also in the central part of the basin and in the south. Lenses of red clastic sediments can be intercalated with the sequence. The lithological equivalent of the Hafik Formation in the west of the basin (the Tuzhisar Formation; Sümengen et al. 1990) has recently been studied sedimentologically (Gündoğan et al. 2005). The Hafik Formation is Oligocene (see below).

Selimiye formation (Kurtman 1973)

This formation is restricted to red clastic sedimentary rocks which have an elongate outcrop in the south of the Sivas Basin. The formation reaches 2100 m in thickness in the Celallı oil exploration well. It comprises alternations of red

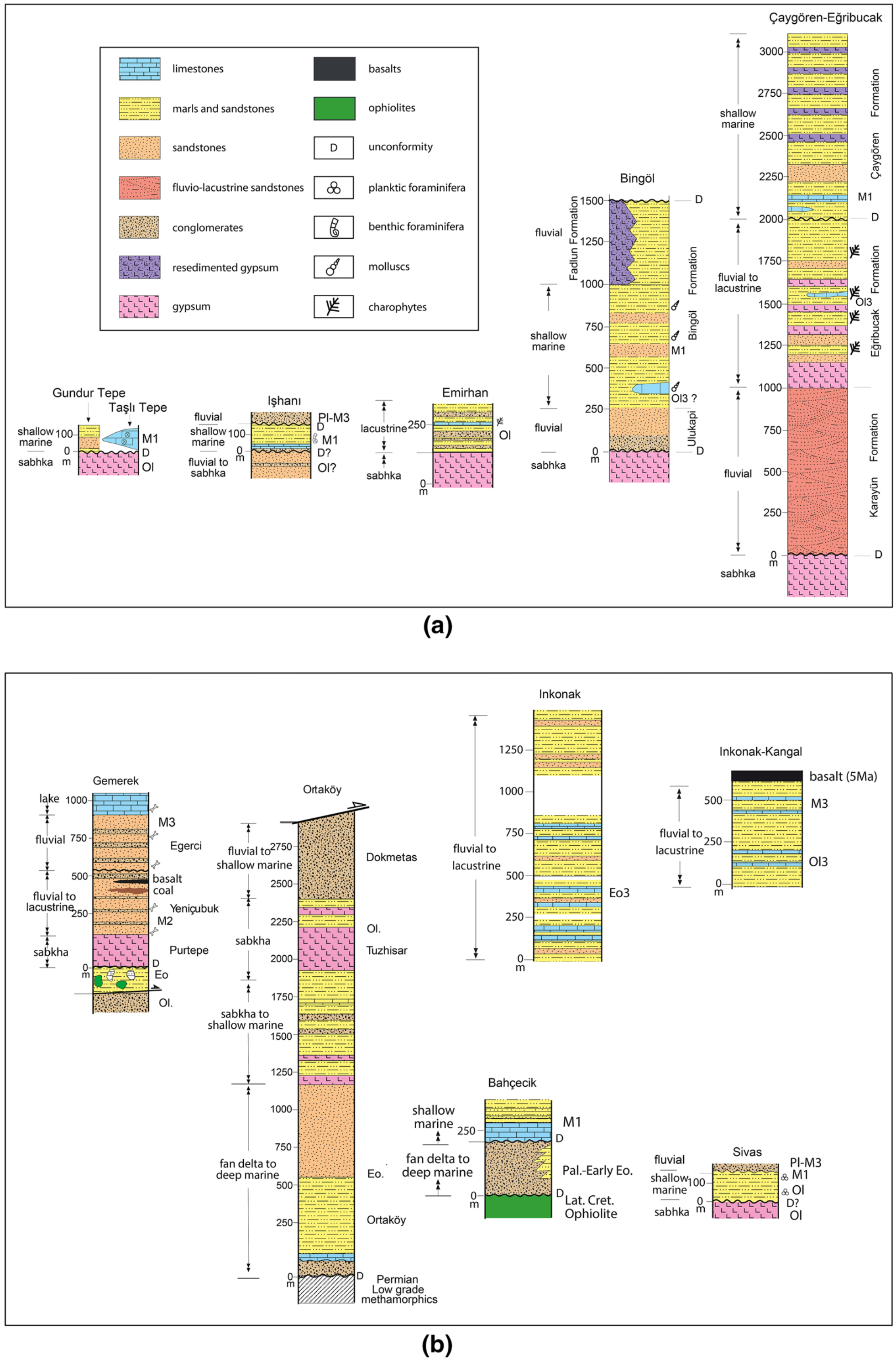

Fig. 3 Logs of local sedimentary sequences in the Sivas Basin. A composite log of the Sivas Basin is not possible because of marked local facies variations. **a** The logs correspond to the area located to the SE of Sivas. Gündür Tepe, Uzuntepe and Işhanı are the sites of open-marine thin sequences with coralgal reefs; Emirhan (old Emirhan) is the site of Oligocene lacustrine sequence; Bingöl and Çaygören-Eğribucak (composite log) are the sites of relatively thick shallow-marine to restricted-marine sequences. **b** In the western part of the Sivas Basin, the logs of Gemerek and Ortaköy, correspond, respectively, to the northern and southern margins of the basin; the other sites are in the central part of the basin, Inkonak near the southern margin, whereas Sivas and Bahçecik are along the northern margin

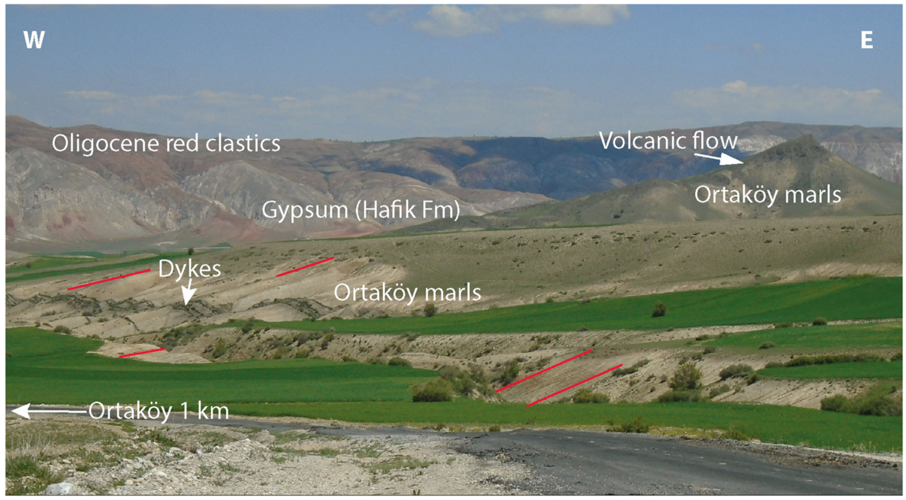

Fig. 4 Intrusive basaltic dykes in the area of Ortaköy (Karaziyaret volcanics; Gündoğan et al. 2005)

Ortaköy volcanism K-Ar age					
Sample	**K* (wt. %)**	**Weight molten (g)**	**40Ar* (%)**	**40Ar* (10-11 moles/g)**	**Age (±2σ)Ma**
2011-09	**1.735 ± 0.017**	**0.51408**	**71.415**	**11.279 ± 0.057**	**37.22 ± 0.42**

Biostratigraphic data from Ortaköy marls below volcanic flow		
Sample	**Planktic foraminifera**	**Age**
2011-38	***Acarinina spinuloinflata, A. bullbrooki, A. collactea, Globigerina venezuelana, Truncorotaloides topilensis, Globigerinatheka subconglobata, G. kugleri, Subbotina hagni, S. eocaena, S. corpulenta, Turborotalia cerroazulensis frontosa, Catapsidrax unicavus.***	**P12-P14 = Middle Eocene (Late Lutetian to Bartonian)**

mudstones, sandstones and conglomerates, with intercalations of anhydrite, gypsum and halite. It has been attributed to an Oligocene age on the basis of microfauna including ostracods (Gökçen and Kelling 1985) and planktic foraminifera (Kavak 1998). However, the environment of deposition of this formation is disputed: it could be marine or instead continental (assuming the marine microfauna is reworked) (Ribes et al. 2015). A continental origin is preferred by us (see Fig. 11); however, the stratigraphy of this formation needs to be reinvestigated.

Karayün formation (Cater et al. 1991)

The type locality of this formation is near the village of Karayün. Three members have been described: the Lower and Upper Karayün sandstone bodies, separated by the Middle Karayün sandstone body. The second member is inhomogenous as it includes lacustrine limestones (with charophytes) and sabkha-type gypsum layers. There are also sub-equal quantities of sandstone and conglomerate, together with subordinate amounts of shale and claystone; a red colour dominates. In the central part of the Sivas Basin, in the area of diapirs (Fig. 2), several minibasins have recently been identified. The infill of these minibasins between the salt diapirs has been interpreted in terms of halokinetic sequences (Ribes et al. 2015).

The Karayün Formation was first assumed to be early Miocene (Cater et al. 1991). However, charophytes within the lacustrine limestones are of late Oligocene age (Poisson et al. 1996, 2012). The Karayün Formation is unconformably overlain by early Miocene marine deposits of the Karacaören Formation.

Karacaören formation (Kurtman 1973)

This formation is synonymous with the Bahçecik Formation of Gökten and Kelling (1991), with part of the Ağılkaya Formation (Çubuk 1994), and also with the Karacaören Group of Poisson et al. (1996). The Karacaören Formation resulted from a renewed marine flooding of the basin. The main part of the Sivas Basin was transgressed, although its western extremity remained lacustrine. The transgressive facies unconformably covered the previous deposits and structures (e.g. near Bahçecik and north of Zara), namely the Hafik gypsum, the red clastic formations (Karayün and Eskiboğazkesen Formations) and also emergent diapirs. The transgressive sediments also overlapped the Central Anatolian Thrust Belt and emergent northerly areas including ophiolites and associated sedimentary sequences derived from the Northern Neotethys. The facies and thickness vary considerably from place to place. Thin sequences are located along the northern margin of the basin (from Sivas to Ishanı, Hafik and Zara; Fig. 3). In contrast, thicker sequences are present in the central part of the basin. Reflecting the major local variation of facies, the formation had been subdivided into five separate members: the Sivas marls; reefs and algal limestones (at Gündür tepe, Uzuntepe (previous name: Taşlı tepe) and Ishanı); the Ulukapı clastics; the Bingöl marls and sandstones; and the Fadlun dere resedimented gypsum. Only the Sivas marls and the coralgal reefs will be discussed here because study of the other members is still incomplete.

The mid- to late Miocene and Pliocene continental formations making up the uppermost levels of the Sivas Basin (which crop out around the Karacaören Formation) will not be considered in detail here (Benlikaya, Incesu and Merakom Formations).

New stratigraphic data

Despite the existing abundant biostratigraphic and isotopic data, many chronological problems remained unsolved or under discussion. We present below significant new chronological data concerning some of the marine Miocene sequences. Previous stratigraphic studies, as cited above, provided good descriptions, mostly of the benthic macrofauna and microfauna within the Sivas Basin. However, these organisms were not widely distributed throughout the basin and they are generally of little interest for precise dating. Isotopic dating mainly concerns the mid-Miocene basalts which are interstratified with continental deposits. For the Gemerek area, these age data are in good agreement with the biostratigraphy based on mammals (Sümengen et al. 1990; Parlak et al. 2001). The K/Ar dating presented here (Fig. 4) is also in good agreement with the data from planktic foraminifera. The new radiometric dating concerns basaltic dykes which cut the Eocene Ortaköy marls (equivalent of the Bozbel flysch) which are generally well dated due to the presence of the classical foraminiferal markers. In contrast, the early Miocene marine sequences are more difficult to date because of the lack of classical stratigraphic markers and also because of reworking of Oligocene species into early Miocene deposits. Of numerous sites which we studied only some provide conclusive biostratigraphic data, as follows: (1) the Sivas marls above the massive gypsum; (2) the Sivas marls above the Uzuntepe coralgal reef; (3) the Çaygören sections. The main new microfossil data, for planktic foraminifera and nannoplankton, are presented in Figs. 7 and 8.

1. Sivas marls around Sivas town (northern slope of the Kızılırmak valley). The location of the sites (Sivas-W and Sivas-EK) is shown in Figs. 2 and 10, and a list of fossils is given in Figs. 7 and 8.

In this area, the Sivas marls, the lowermost marine transgressive deposits, directly cover the massive gypsum of the Hafik Formation, which is Oligocene in age. Poisson et al. (1997) first described this formation and attributed it to the Oligocene and Miocene on the basis of planktic foraminifera. Following new studies based on nannoplankton and also a revision of the planktic foraminifera, an Oligocene age can now be excluded and it is inferred that the base of the formation is Aquitanian. Oligocene planktic foraminifera in the assemblages are now interpreted as being reworked into Lower Miocene marls. The highest levels of the formation are now dated as early Langhian.

The main facies of the Sivas marls are grey marls, reddish sandstones and subordinate conglomerates. Although the thickness is difficult to determine exactly owing to the multiple outcrops, it may reach 100–200 m. The source of the marine flooding could not have been from the west or the south-west, where lacustrine sequences were deposited at the same time. An origin from the east and south-east was inferred by Erünal-Erentöz (1956). However, the existence of a similar marine sequence to the north of Zara (100 km east of Sivas and 40 km north of Zara) shows that a northern connection with the Black Sea Basin cannot be completely excluded.

The base of the Sivas marls is well exposed along the road to Ankara (Fig. 10), between the last buildings in the town and the « cimento fabrikası » (Sivas-W site; Figs. 2, 5, 7, 8, 10). The basal beds above the gypsum are composed of 5–10 m of reddish, fine-grained sandstones with plant remains, surrounded by green to grey marls. The nannoplankton assemblages are Aquitanian in age. Pectinids (pelecypod) shells are the dominant macrofauna in these marls, including *Oopecten rotundata*, which is a good marker of the Aquitanian in the Central Paratethys Basins (Mandic 2007). The microfauna, predominantly foraminifera, are rich and abundant but remain poorly determined because of the lack of the classical markers. However, Aquitanian, Burdigalian and Langhian ages have been tentatively identified on the basis of secondary markers. The presence of marine Langhian deposits is thus established for the first time in the Sivas Basin.

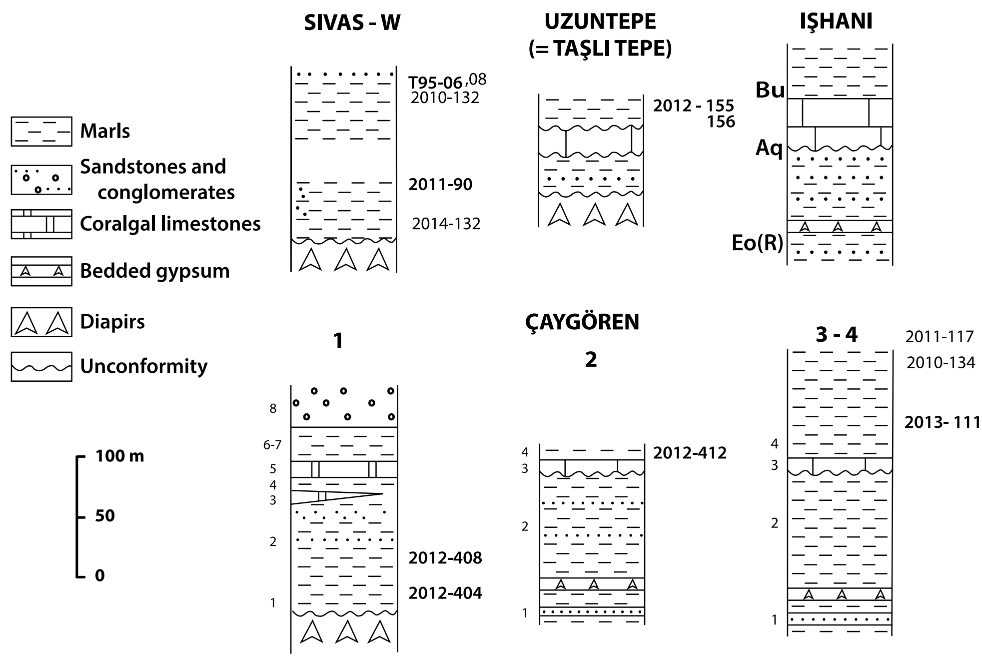

Fig. 5 Logs of the sites described in the text

2. Uzuntepe coralgal reef

The main coralgal reef is located to the SE of Sivas on Uzuntepe hill (also known as Taşlıtepe on the ancient maps), near the left (southern) bank of the Kızılırmak River (see Figs. 2, 5, 7, 8). In this area, massive limestones emerge from the alluvial plain of the Kızılırmak River to form an isolated hill. The limestones are composed of large in situ coral colonies, more or less encrusted by colonies of red algae. The southern slopes of the hill are covered by scree and slope deposits (following the recent abandonment of small quarries along the slope), such that the substratum of the massive limestones cannot be easily observed. In some places (between houses in Uzuntepe village), the presence of some outcrops of red clastic beds and gypsum suggests that the substratum is represented by the Hafik Formation in the form of massive gypsum (a possible salt diapir). A locally developed hardground, encrusted by iron oxide, caps the limestones which are surrounded by the Sivas marls. These marls include thin-shelled pelecypods and rich assemblages of foraminifera. Ostracods are present but not abundant. Echinoid bioclasts occur frequently. At other sites in the basin several smaller patch reefs, generally associated with red algae, have been observed locally at the base of the marine deposits (north of Hafik; in the Bingöl basin near Sivas; in the Çaygören area (see below); and elsewhere in the Sivas Basin.

The algal limestones resulted from an accumulation of more or less in situ fragments of red algae (Melobesiae), the thickness of which can reach 50 m. Some beds contain early Miocene benthic foraminifera (*Miogypsina*; *Amphistegina*). The algal limestones grade upwards into grey marls rich in molluscs and echinoderms, as first studied near Ishanı (SE of Sivas) (Chaput 1936). Associated corals are widely distributed throughout the basin, where they underline the base of the marine transgressive sequence. The westernmost Miocene marine deposits in the Sivas Basin, dominated by accumulations of ostrea, crop out near Gücük (near Şarkışla) and to the south of Altınyayla.

3. Çaygören sections (location Figs. 2, 6; fossils Figs. 5, 7, 8)

The sites are located around the village of Çaygören, along the east side of the valley (Figs. 2, 6; see also Figs. 5, 7, 8 for fossil illustrations). The base of the sequence overlies a diapir. Both the diapir and the overlying Miocene sequence were folded and thrust northwestward during the late Miocene compressional phase (Fig. 10). The sections described below illustrate the variability of the Miocene basal transgressive facies in the central parts of the Sivas Basin, above salt diapirs. It is notable that the basal clastics do not contain fragments reworked from the underlying salt diapirs. In contrast, pebbles and gypsarenite are abundant in the Fadlun-type sequence which surrounds the marine beds and the lower continental beds (see Fig. 6).

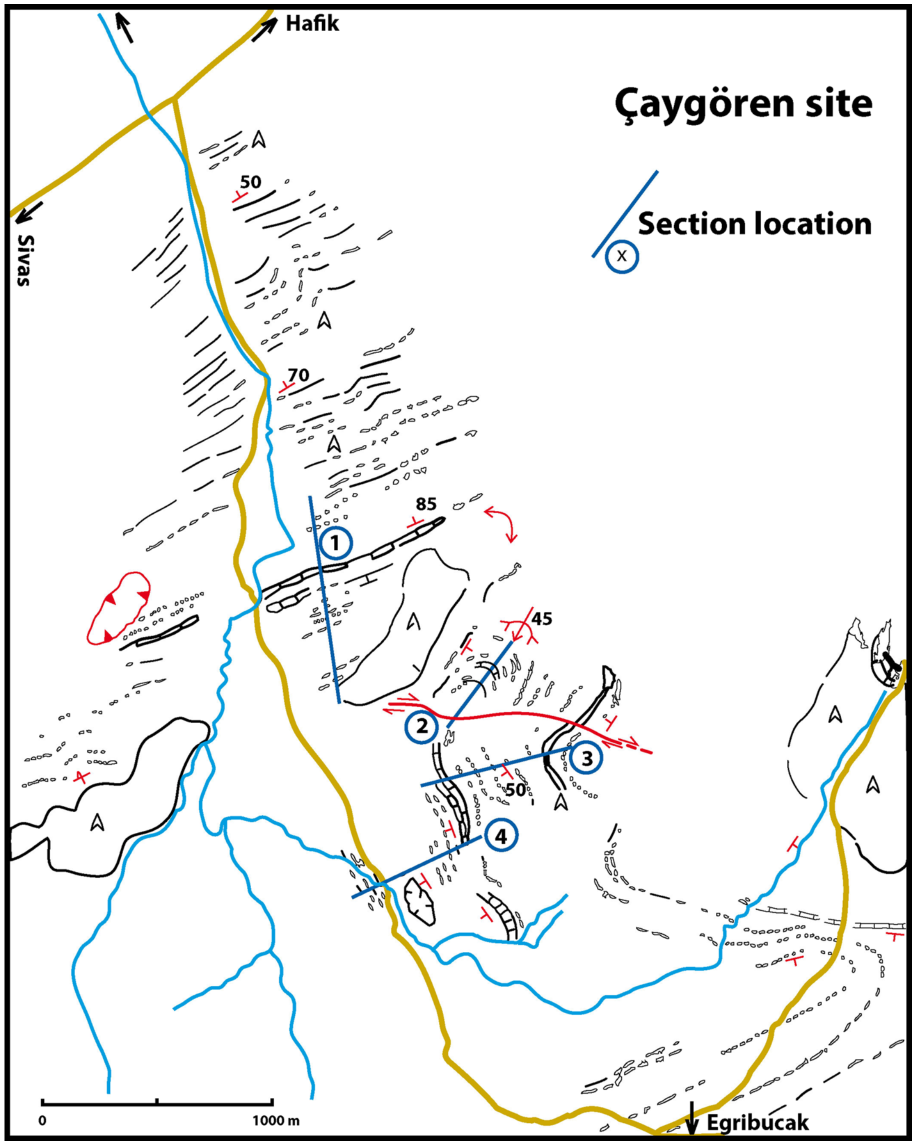

Fig. 6 Schematic geological map of the Çaygören area showing the location of the sections described in the text

In section No. 1 (Figs. 5, 6), the sequence is composed, from the base to the top, of the following succession:

1. Fine-grained green marls with rich assemblages of planktic foraminifera and nannoplankton, the latter being dated as Aquitanian.
2. Graded marly sandstones including various bioclasts of molluscs, echinids and rare fish otolites. Nannoplankton again indicate an Aquitanian age.
3. An initial lens (50 m long by 5 m maximum thickness) of massive bioclastic limestones is composed of fragments of molluscs, corals, red algae and various benthic foraminifera.
4. Further marls and sandstones include a bed composed of white pelecypods shells. Such lumachelle have been frequently observed at other sites in the Sivas Basin.
5. A second bed of bioclastic limestones (several metres thick) has a lenticular geometry (each lens 100–200 m long, but gives a deceptive appearance of a continuous bed). Some of these limestone lenses are dominated by small coral colonies. The colonies are disaggregated, but the fragments (decimetric in size) have not been rounded to form pebbles, suggesting that they are more or less in situ. Some other beds are restricted to fragments of red algae, more or less reworked.
9. Several sandstone beds occur within the bioclastic limestones. These are characterised by abundant small benthic foraminifera such as miliolids and peneroplids and fragments of red algae (Sirel et al. 2013).
10. Sandy marls and sandstones surrounding limestones. These include reworked assemblages of foraminifera and fragmented molluscs and grade into a red terrigenous clastic sequence in the form of conglomerates,

Fig. 7 Illustrations of the species of planktic foraminifera. In the absence of the classical markers, other species have been used, tentatively, for stratigraphic determination: Sivas-W:2014-132 is Lower Miocene; T9508, 2011-90 and 2010-132 are attributed to the late Burdigalian-early Langhian due to the presence of the species *foliata*, *lentiana*, *eamesi*. The presence of *C. chipolensis* excludes an age younger than middle Langhian. The Uzuntepe marls could be late Burdigalian (*foliata* and *eamesi*). In the Çaygören sections, despite abundant microfauna rich in planktic foraminifera, no markers have been found. 2013-111 at the base is Aquitanian, and 2011-117 and 2010-134 are late Burdigalian-early Langhian (equivalent of the Karpatian of Paratethyan basins)

	SIVAS – West					Uzun tepe		CAYGÖREN		
	2010 - 132	T95 - 06	2011 - 90	T95 - 08	2014-132	2012-155	2012-156	2011 - 117	2010 - 134	2013 - 111
Globigerinoides bolli										
G. quadrilobatus	+				+					
G. trilobus		+		+						
G. bisphericus									+	
G. altiaperturus		+			+					
G. sacculifer					+					
G. apertasuturalis								+		
Praeorbulina suturalis		?								
P. transitoria								aff	?	
Globigerina foliata	+		+	+		+	+	+	+	
G. praebulloides	+	+	+	+		+ ab	+	+	+	+
G. bulloides					+	aff		+	+	
G. venezuelana										aff
G. falconensis						aff				
G. juvenilis			+						+	+
G. lentiana						+	?	+		
G. officinalis	+						?		+	+
G. occlusa									+	
G. concinna						aff			+	
G. globularis						+		+		
G. eamesi							+		?	
G. ottnangiensis								+		
G. falconarae								+		
G. tarchanensis			+							
Dentoglobigerina larmeui					+					
D. baroemoenensis										
D. galavisi									aff	aff
Tenuitella clemenciae	+		gr					+	+	+
T. gemma	+								+	+
T. angustiumbilicata	+							+		+
T. pseudoedita								+		
Globigerinella obesa	+		+					+	+	+
G. pseudobesa										
G. praesiphonifera			+							
Globoturborotalita druryi				aff				+	aff	
G. woodi woodi				+						aff
G. parawoodi				+	+					
G. pseudociperoensis										
G. ottnangiensis										
G. angulisuturalis										+
G. anguliofficinalis	?									+
G. ciperoensis	+									+
Turborotalita quinqueloba	+							+	+	+
Globoturborotalia connecta	+							+		
Cassigerinella chipolensis			+			+	+ ab	+	+	+
Globorotaloides permicrus										+
G. suteri								aff		
G. stainforthi					+					
Paragloborotalia nana						+				+
P. continuosa								+		
P. pseudokugleri										
G. uvula-juvenilis gr.								+	+	
Globoquadrina altispira										
G. dehiscens					+					
Catapsydrax martini									+	+
Globigerinita incognita								?		
Beella clavacella	aff		+					aff	aff	
Bolboforma reticulata			+							

sandstones and red pelitic layers which are probably of continental origin.

11. The uppermost part of the succession is composed of a several-hundred-metres-thick conglomeratic sequence, which is correlated with the Fadlun Formation (sensu Poisson et al. 1996; work in progress). There are about 40 layers of conglomerates (one metre to several metres thick each) that are exclusively composed of gypsum pebbles, interbedded with layers of gypsarenites and marls.

Fig. 8 Listing of the nannoplankton species

	Cyclicargolithus floridanus	*Cyclicargolithus abisectus*	*Sphenolithus moriformis*	*Helicosphaera euphratis*	*Pontosphaera spp.*	*Helicosphaera carteri*	*Coccolithus pelagicus*	*Cyclococcolithus formosus*	*Dictyococcites spp.*	*Discoaster deflandrei*	*Helicosphaera spp.*	*D. bisectus*	*S. disbelemnas*	*BIOSTRATIGRAPHY (Fornaciari et al. 1996, Raffi et al. 2006)*
SIVAS West														
2012-155	C	C	R	R	R	A	A							MNN1-2
2012-156	C			R	C	A	C	RR	R	R	R			MNN1-2
2012-59	Barren													
2012-146	very rare nannos Late Eocene-early Oligocene													all reworked?
2012-141	very rare nannos Late Eocene-early Oligocene													
2012-54	very rare nannos Late Eocene-early Oligocene													
2012-52	very rare nannos Late Eocene-early Oligocene													
2012-142	very rare nannos Late Eocene-early Oligocene													
2014-145	almost barren													
2014-143	R												R	MNN1-2
2014-141	C												R	MNN1-2
2014-82	A	C	R	R	R	C	A							MNN1-2
2014-80	A	C	R	R	R	R	A							MNN1-2
2014-109	C	C	R	R	R	A	A							MNN1-2
2014-78	C	C	R	R	R	A	A							MNN1-2
SIVAS East														
2014-174	C	C	R	R	R	A	A							MNN1-2
2014-173	C	C	R	R	R	A	A							MNN1-2
BAKIMLI														
2014-90	barren													
2014-91	R	R					R							MNN1-2
2014-92	C	C	R	R			C						R	MNN1-2
2014-93	C	C	R	R	R		C						R	MNN1-2
2014-94	C	C	R				R							MNN1-2
2014-95	C	C	R				R							MNN1-2
2014-96	R						R							MNN1-2
2014-97	C	C	C	C	R		C						R	MNN1-2
TUZLAGŐZŰ														
2014-188	R												R	MNN1-2
2014-184	A	A	R	C	R	R	C							MNN1-2
2014-183	A	A	R	R		R	C							MNN1-2
ÇAYGÖREN														
404	A	C	R	R	R	A	A							MNN1-2
408	C			C	C		C	RR	R	R				MNN1-2
412	C			C	C		C	RR	R	R				MNN1-2

In sections No. 2, 3, 4 (Figs. 5, 6), located near the section No. 1, on the opposite side of the diapir, the sequence is rather different. From base to top it comprises:

1. Coarse conglomerates (2–50 m thick) which include extra-basinal pebbles (quartzites, ophiolites, limestones).

2. Marls and sandy limestones with rich assemblages of benthic foraminifera (miliolids and peneroplids; Sirel et al. 2013), numerous ostracods and various bioclasts (echinids; molluscs), and locally, charophyte remains. The depositional environment was a more or less restricted shallow sea (not open-marine).
3. Massive algal limestones form lenses interbedded with lenticular coral colonies. The red algal and coral colonies are disaggregated but more or less in situ. Large benthic foraminifera (miogypsinidae), attributed to the Lower Miocene, are commonly associated with red algae but not with miliolid–peneroplid limestones. In the Sivas Basin, these coralgal limestones constitute a key horizon that is generally located near the base of the open-marine sequence. The coralgal limestones in the Çaygören area represent the lateral equivalent of the Uzuntepe coralgal reef. However, they are not present around Sivas at the base of the marls.
4. Green marls (c. 200 m thick) are interbedded with rare beds of fine-grained sandstone. These marls contain rich assemblages of planktic foraminifera, echinoid remains and rare ostracods. At the base, these marls are dated as Aquitanian utilising nannoplankton assemblages. The top of the succession is dated as Burdigalian-lower Langhian based on planktic foraminiferal assemblages. Their deposition corresponds to an important environmental change within the Sivas Basin from restricted-marine during the deposition of the marls below the coralgal limestones, to more open-marine during the deposition of the coralgal limestones, and then deeper marine when the marls above the limestones were deposited.

Biostratigraphy

In the Sivas Basin, the P–B (planktic/benthic) ratio is generally very low, typically from 20 to 40 %. In the Çaygören area, the ratio reaches 90 % in a single sample. This high ratio could indicate a shallow-marine environment or a restricted, poorly oxygenated environment, with a sporadic influx of surface marine water from the open sea above a varying threshold. Such a setting is consistent with the observed assemblage of benthic foraminifera, which is dominated by a single species (*Nonion boueanum*) that is indicative of sub-oxic conditions. By contrast, the assemblage of benthic foraminifera in the Sivas marls is characteristic of an open-marine environment with *Oridorsalis umbonatus*, *Textularia* sp., *Melonis affinis*, *Martinottiella communis*, *Cassidulina margareta* and rare ostracods (e.g. *Krithe).*

Planktic foraminifera are abundant in the Sivas marls and in the marls above the coralgal limestones in the Uzuntepe and Çaygören sections. In the Çaygören sections, an additional sequence below the coralgal limestones marls lack planktic foraminifera. However, these marls contain abundant assemblages of small benthic foraminifera (e.g. Miliolidae and Peneroplidae; Sirel et al. 2013) and also ostracods that are characteristic of a shallow-water environment. A gypsum bed occurs near the base of the sequence.

In the Çaygören sections, above the coralgal limestones, the planktic foraminiferal assemblages are distinguished by a relatively rich fauna, composed exclusively of small-sized individuals of low species diversity. Just above the coralgal limestones (samples 2014-111), the dominant taxa are *Globigerina praebulloides*, *G.* sp., *Globigerinella obesa* and Tenuitellids that all are long-ranging Oligo-Miocene species. However, nannoplankton in the same beds indicate an early Miocene age (late Aquitanian). The uppermost beds (samples 2010-134; 2011-117) of the marls contain rare specimens of *Globigerina lentiana*, *G. foliata* and *G. ottnangiensis*, and also Tenuitellina, which is suggestive of a late Burdigalian age, possibly as young as the Karpatian-early Badenian (Langhian). Despite this, the ubiquitous occurrence of *Cassigerinella chipolensis* indicates an age not younger than middle Langhian. These unusual assemblages lack the classical Miocene biomarkers, as known in the Mediterranean area, including *Globigerinoides*, *Globoquadrina*, *Catapsydrax*, *Globorotalia*, *Paragloborotalia* (e.g. *P. kugleri* group) and *Dentoglobigerina*.

For the study of the nannoplankton, smear slides were prepared directly from sediment samples and the nannofossil assemblages were then analysed using a polarising light microscope at 1250× magnification. The total abundance of nannofossils was estimated by comparing their occurrence with other biogenic and inorganic components. The species abundance was reported semiquantitatively after estimating the relative abundances (see Fig. 8).

Where the environment was restricted, marine samples are devoid of nannoplankton or only contain rare specimens of non-age-diagnostic long-ranging species, generally of Eocene–Oligocene age. In contrast, the assemblages are more abundant and better preserved in the open-marine environments. In the sections around Sivas and Çaygören, the relatively high abundance of *H. carteri* compared to *H. euphratis* indicates that the samples come from above the *H. carteri/euphratis* crossover (see Raffi et al. 2006; Fornaciari and Rio 1996; Fornaciari et al. 1996), of late Aquitanian age (Fig. 9). The assemblages can be generally assigned to the MNN 1-2 zone of Fornaciari et al. (1996). In addition, the species *S. disbelemnos* is characteristic of the early Aquitanian. However, this species was rarely

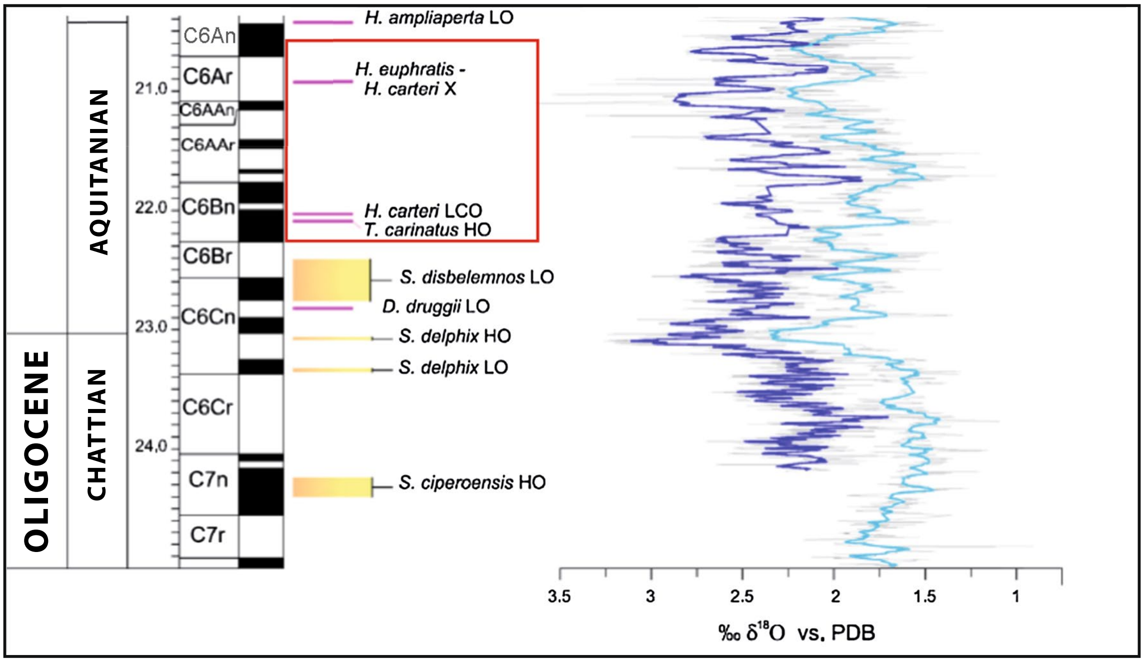

Fig. 9 Chronostratigraphic scale for the nannoplankton (modified after Raffi et al. 2006) with a summary of the positions of the biohorizons in the time interval 14–25 Ma, relative to the (delta O18) records of ODP Sites 1146 (*orange line*) and 1237 (*ochre line*; Hollburn et al. in press), Site 925 (*green line*; Shackleton 2001), Site 1090 (*dark blue line*; Billups et al. 2004 and Sites 926/929 (*sky blue line*; Zachos et al. 2001). Per mil values are corrected for sea-water equilibrium, using a correction of +O.64 (Shackleton and Hall 1997). Biohorizon positions are relative to the polarity timescale with magnetic reversal and stage boundary ages based on ATNTS-2004 (Lourens et al. 2004). Lüdecke et al. (2013) for the Turkish Central Anatolian Basins

observed and further division into early or late Aquitanian cannot be made in most parts of the sites studied.

The data given above indicate an Aquitanian age for the marine transgression in the Sivas Basin, at least for the central-northern part of this basin, around Sivas.

Discussion: Towards a Cenozoic palaeogeographic reconstruction of Eastern Anatolia

There are several questions concerning the origin and evolution of the Sivas Basin: (1) the nature of the basement and the timing of ophiolite emplacement; (2) the salt tectonics and thrust tectonics; (3) the age of the gypsum sequences; (4) the Cenozoic palaeogeographic evolution of the basin.

Basement of the Sivas Basin

There are currently two main models: (1) the Kırşehir Massif and the Taurus were separated by a Mesozoic oceanic basin called the Inner Tauride ocean. (2) The Kırşehir Massif formed a part of the Taurus belt, and thus, the substratum of the Sivas Basin was the Taurus.

The first model, that of an Inner Taurus Basin, was first proposed by Sengör and Yilmaz (1981) who located it between the Pütürge-Bitlis Massifs and the Munzur Dağ (easternmost Taurus) (i.e. to the south of the central and eastern Taurus), with a possible extension to the SW in the direction of the Pamphylian Basin (possibly through the recently identified Berit oceanic basin; Robertson et al. 2006). Görür et al. (1984, 1991) considered that the CACC and the Taurides developed differently, separated by the Inner Tauride basin. The Inner Tauride basin was displaced from its original position such that it is now located to the west of the CACC and between the Taurus belt and the CACC (i.e. to the north of the central and eastern Taurus). Although different from the initial model of the Intra Tauride basin of Şengör and Yilmaz (1981), this model has nevertheless been widely accepted. The Intra Tauride basin was adopted in the maps resulting from the Tethys and Peritethys Programmes (Dercourt et al. 1993, 2000; Barrier and Vrielynck 2008) and in other publications (Şengör 1984; Görür et al. 1984, 1991; Gökten 1993; Temiz 1994, 1996; Temiz et al. 1993; Koçyiğit and Beyhan 1998; Robertson et al. 2013; Oberhänsli et al. 2010; Pourteau et al. 2010). Booth et al. (2014) locate the Inner Tauride ocean to

Fig. 10 *1* In situ colony of corals (*Calamophylia*?); *2* *Oopecten rotundatus* from *3*, showing the base of the Sivas marls; *4* top of the Sivas marls; *5–6* Sivas thrust (late Miocene). View towards the east from the road to Ankara; *7–8* folds and thrusts in the Çaygören area. The Çaygören diapir and its lower Miocene cover are folded and thrusted towards the NW in relation with the late Miocene compression which initiated the Sivas thrust (see also the map in Fig. 6 where the syncline is shown; Sect. 2)

the north of the Tauride–Anatolide platform and viewed it as a supra-subduction oceanic basin that was created to the south of the Izmir–Ankara–Erzincan Ocean.

One important argument in favour of the Intra Tauride basin model relates to the existence of ophiolites in the Sivas Basin below the Tertiary infill. Ophiolite outcrops occur from place to place in the central part of the basin, and the oldest beds of the sedimentary fill were deposited on the top of the ophiolites, as seen in the south (Artan and Sestini 1971) and also in the western part of the basin, near Bünyan (Dirik et al. 1999). In the central part of the basin, ophiolites have been exhumed related to faulting, for example, related to the Deliler Fault. Along the southern part of the basin an important slice of ophiolites is intercalated with other tectonic slices in the Tecer Dağ. Ophiolites are also present on the Tauride platforms, eastwards as far as Divriği and the Kemah-Erzincan areas. In this model, these ophiolites (mostly peridotites, serpentinites and ophiolitic mélanges) are interpreted as remnants of the oceanic crust of the Inner Tauride ocean. Booth et al. (2014) illustrate the

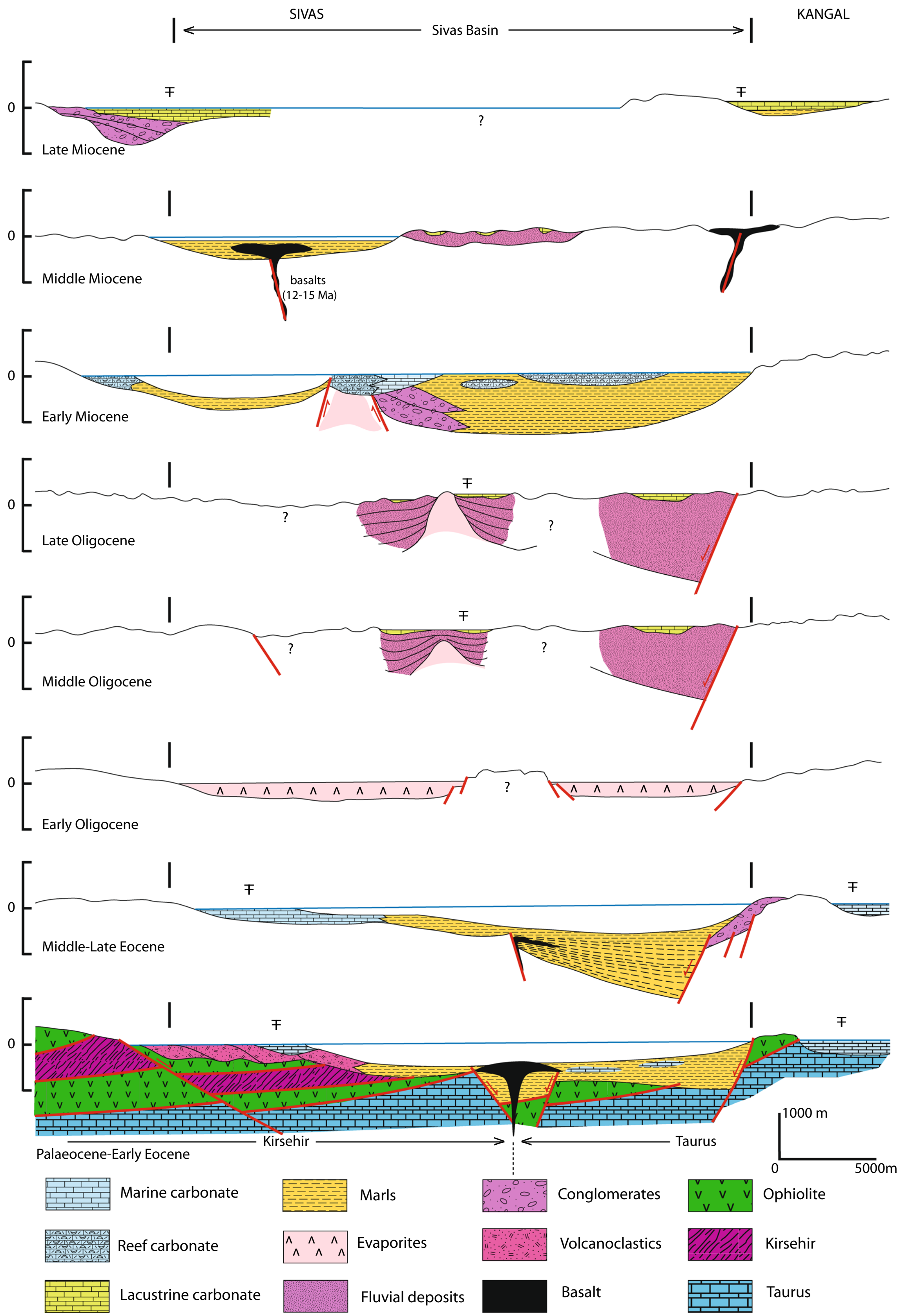

Fig. 11 Cross-sectional interpretations of the structural evolution of the Sivas Basin. In the Oligocene sections, we postulate that the Selimiye Formation, which separates the central minibasins from the Taurus, was deposited in a continental environment (see text for discussion)

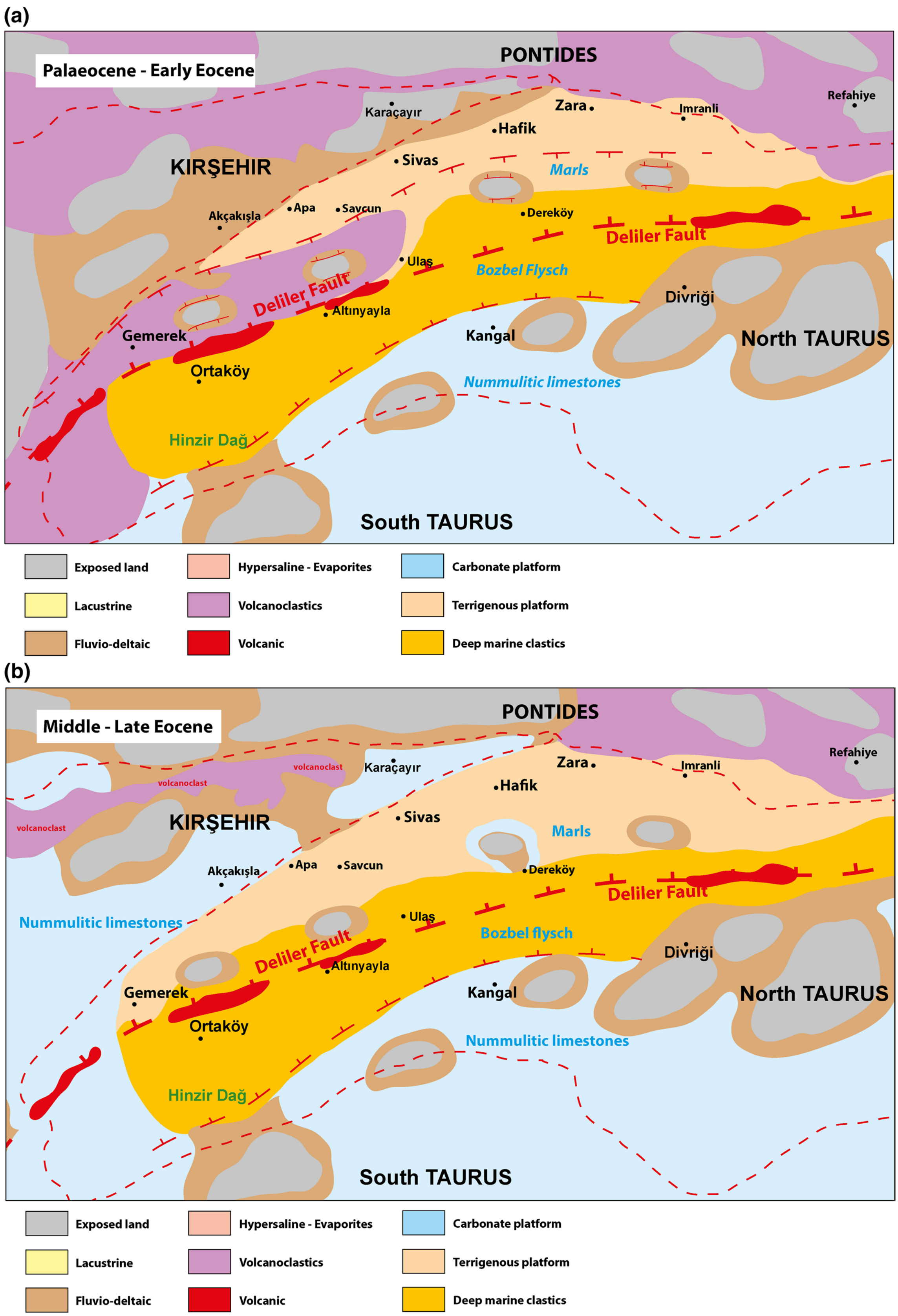

Fig. 12 Palaeogeographic maps from Palaeocene to late Miocene times (see text for discussion)

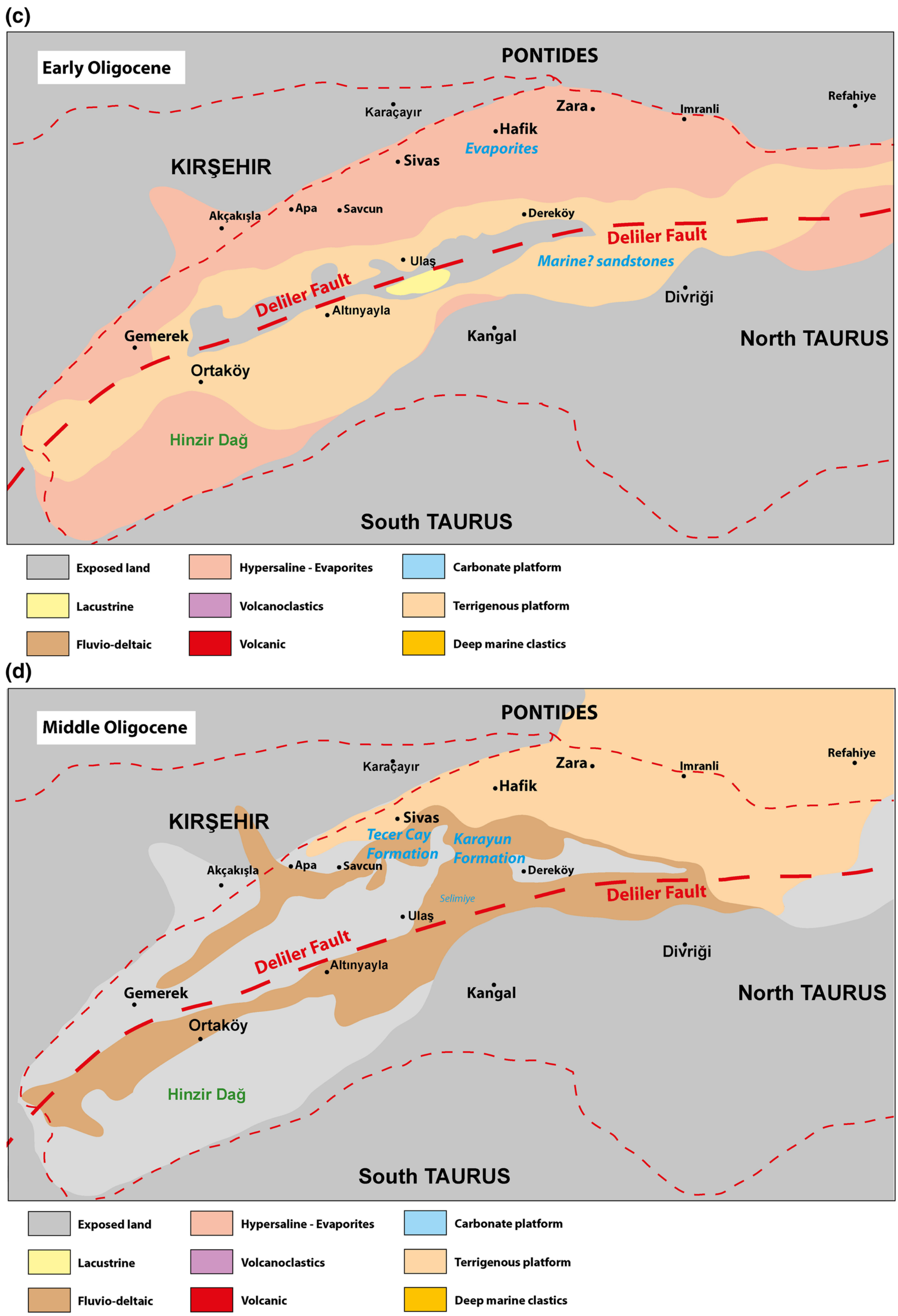

Fig. 12 continued

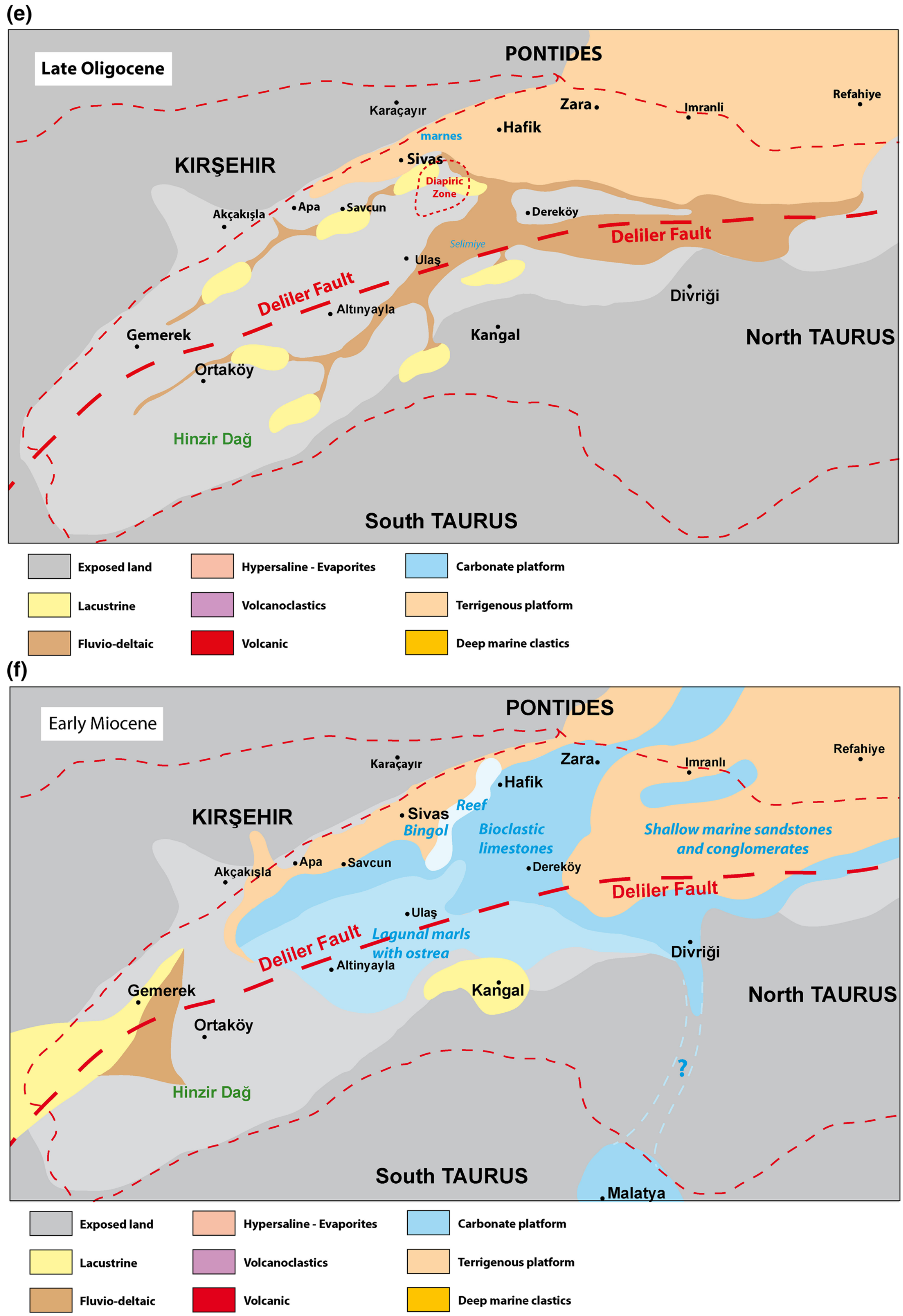

Fig. 12 continued

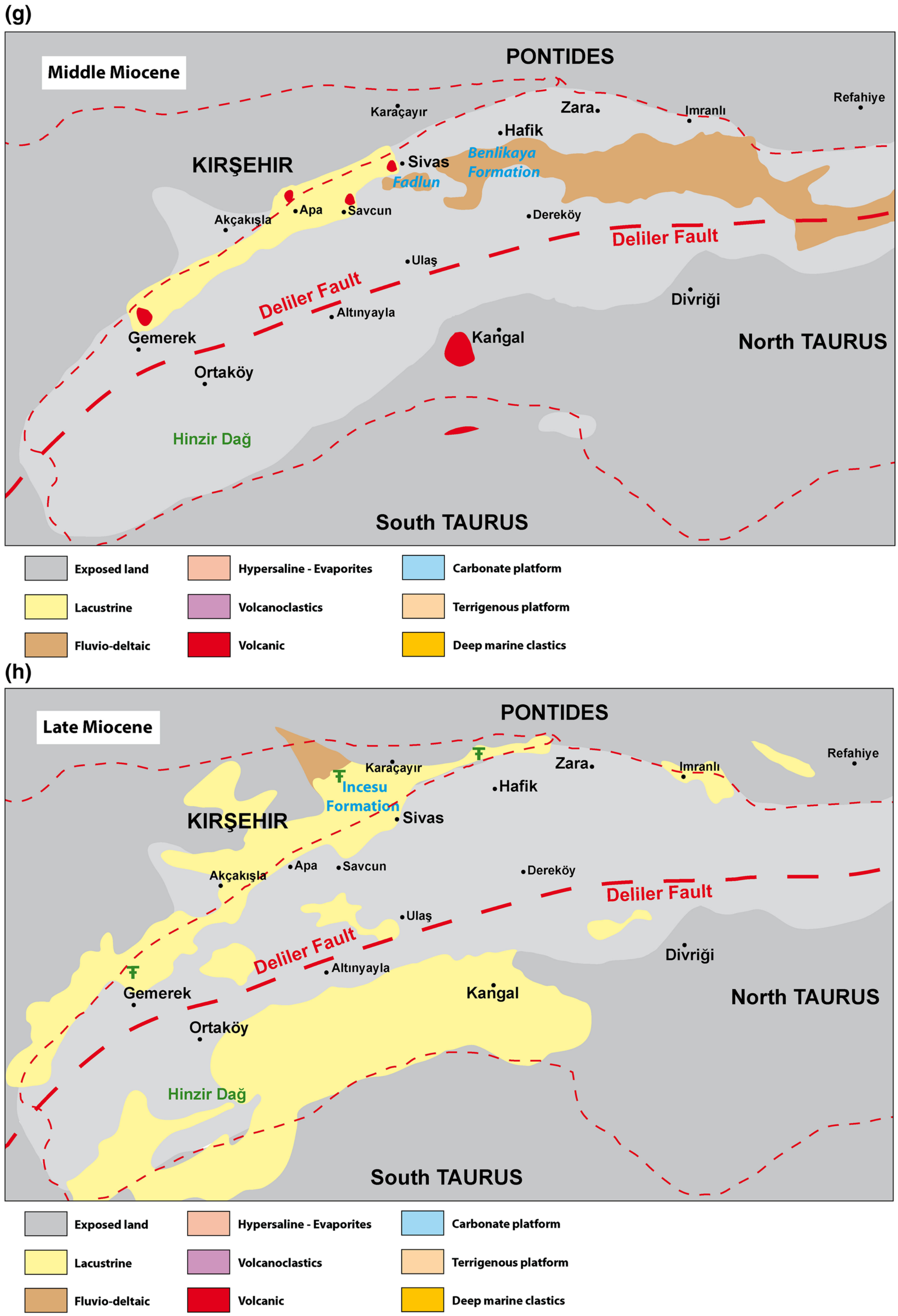

Fig. 12 continued

supra-ophiolite basin model in their study of the Hekimhan area. This model also provides a solution for the problem of the position of the poorly known Kırşehir Block, which was viewed as an isolated block, reflecting its uncertain palaeogeographic setting.

In the second model, all of the ophiolites originated from the Northern Neotethys (Izmir–Ankara–Erzincan suture zone), from which they were obducted and transported over and above the Kırşehir Massif towards the Taurus belt during late Cretaceous times. These highly allochthonous ophiolites include the Central Anatolian Ophiolites (CAO; Yalınız et al. 2000), the Sivas Basin ophiolites and the eastern Tauride ophiolites.

In their model, Moix et al. (2008) considered that the Kırşehir Massif was accreted to the Taurides during the Triassic, and as a consequence, there was no place between them for a Mesozoic Inner Tauride basin. Dirik et al. (1999) discussed the evolution of the westernmost part of the Sivas Basin, near Kayseri. In this area, the Sivas Basin rests on the metamorphic sequences of the southern CACC to the north, but on the recrystallised limestones and clastics of the Permian-lower Cretaceous Bünyan metamorphics to the south. Because the metamorphic soles are almost identical, the authors concluded that there was no evidence for a deep oceanic basin between them. In this area, blueschists occur in pebbles in a late Cretaceous conglomerate that occurs within the lower sequence of the southern margin of the Sivas Basin and also in an Eocene olistostrome. This suggests that they cannot be related to subduction in this area as suggested by Oberhänsli et al. (2010) and Pourteau et al. (2010), but where did they come from?

The timing of the emplacement of the ophiolitic nappes from the north is important to understand the nature of the basement of the Sivas Basin. It is generally agreed that the ophiolitic nappes were emplaced during the late Cretaceous as elsewhere in central and southern Turkey. The timing of the ophiolite emplacement is precisely documented in several sites: (a) on the CACC, around Sarıkaraman (Yalınız et al. 2000), where the Central Anatolian Ophiolites (CAO) were emplaced after the late Santonian but before pre-middle Campanian; (b) similar ages have been inferred for the thrusting of the ophiolitic nappes and ophiolitic mélanges on the eastern Taurus platforms to the south of the Sivas Basin. In the Hekimhan area, for instance, the ophiolitic nappes are directly covered by shallow-marine rudist limestones attributed for a long time to the Campanian (Douvillé 1896 cited in Chaput 1936; Görmüs 1993; Özer 2002; Özer et al. 2008, 2009; Robertson et al. 2013; Booth et al. 2014). A similar relationship is present on the Arabian platform in the region of Adiyaman (Özer et al. 2008); (c) between Gürün and Kangal (around and below the Kangal Neogene basin), the northern part of the Taurus platform consists of massive carbonates, with diagenetic cherts locally. These limestones probably represent the westernmost lateral equivalent of the Northern Allochthon of Robertson et al. (2013). They contain poorly diversified bioclasts of shallow-marine environments which could be attributed to the Mesozoic. Remnants of an ophiolitic mélange are locally preserved on these carbonates. In the village of Avşarören (SW border of the Kangal Basin, near the road from Kangal to Gürün), we have recently observed an ophiolitic mélange resting on top of the Taurus limestones and unconformably covered by limestones and marls including rudists (*Colveraia variabilis* Klinghart, *Balabania acuticostata*, and *B. densicostata* Karacabey-Öztemür) of upper Campanian-lower Maastrichtian age.

As a result, in several places in central Anatolia (CACC) and in the Eastern Taurus, the emplacement of the ophiolitic nappes is synchronous and is dated as pre-late Campanian. Another common point concerns the granitic intrusions. The late Cretaceous–Palaeocene intrusive plutonic bodies are mainly located along the northern margin of the Kırşehir Massif (facing the Neotethyan suture), and along the Taurus platforms, from Divriği to the east, also directly facing the Neotethyan suture.

At a regional scale, we propose the following model: the Kırşehir Massif and the Taurus belt experienced a quite similar geodynamic evolutions at the end of the Cretaceous: (1) they were probably close together without the Inner Tauride basin between them; (2) they were covered at the same time, probably by the same ophiolitic nappe expelled from the Ankara–Erzincan suture zone; (3) they were intruded by similar plutonic rocks (granites), at the same time (late Cretaceous–early Palaeocene after emplacement of the ophiolites which are also intruded by the granites). We, therefore, favour the second model in preference to the hypothesis of an Intra Tauride basin. We interpret the ophiolites as all of supra-subduction zone type, derived, synchronously, from the Northern Neotethys oceanic basin.

Salt tectonics versus thrust tectonics

The existence of salt diapirs in the Sivas Basin has long been known (Chaput 1936; Stchepinsky 1939). More recently, Çubuk (1994) and Çubuk and Inan (1998) described one example of a diapir near Emirhan, while Çiner et al. (2002) gave a map of the diapirs in the same area. However, the role of salt tectonics was neglected in the proposed fold and thrust belt model (Poisson et al. 1996; Guezou et al. 1996). In contrast, a model centred on salt tectonics has been recently proposed, which reduces the role of the thrusting (Ringenbach et al. 2013; Callot et al. 2014). This model mainly relates to a restricted area of the Sivas Basin to the SE of Sivas (Emirhan, Karayün, Çaygören, Kızılkavraz areas). However, the existence of thrusts has been clearly demonstrated, along the northern margin of the Sivas Basin and also in the central part of the basin (Poisson et al. 1992, 1996). The role of thrusting versus salt tectonics is currently under study. As a preliminary conclusion, the Sivas thrust and related thrusts resulted from a late Miocene-Pliocene compressional event. The directions of tectonic transport during this phase are remarkably constant, towards the NW or NNW (based on thrust planes, reverse faults and folds). In contrast, the salt diapirs and the related salt tectonics were mainly active during the Oligocene (or even the lower Miocene). The salt tectonics do not necessitate a compressional event, although compression could have contributed to the exhumation of the diapirs. The preferred orientation of the salt tectonic structures may have not been primary but could have resulted from the late Miocene compression (Figs. 1, 2, 6, 10).

Age of the gypsum deposits

The age of the gypsum has been disputed since 1956 until now. Several key topics of discussion are as follows:

Massive gypsum and layered gypsum There has been confusion between the Hafik Gypsum (lower Oligocene), generally described as "massive gypsum", and the contrasting "layered gypsum" (mainly Miocene), which corresponds to the reworked/remobilised Oligocene gypsum in the Miocene deposits (including conglomerates, gypsarenites, "new" sabkha deposits = Fadlun Dere Formation of Poisson et al. 1997).

Horizontal transport This resulted from the final compression generating duplications of the sequence, with the gypsum layers acting as a preferential décollement level. Because these duplications were not properly recognised, the tectonic position of the Oligocene gypsum above the early Miocene was thought to reflect a transitional normal contact with gypsum of Miocene age, despite several earlier publications which established its Oligocene age.

Stratigraphic data Stchepinsky (1939) was among the first to attribute an Oligocene age to the gypsum. This age was confirmed by Lahn in 1950 and again in 1957. In these works, the Oligocene age was inferred from the position of the gypsum between well-dated formations: i.e. late Eocene (below) and lower Miocene (above). This, however, was later ignored such that Nebert (1956) retained a middle and late Miocene age, while Kurtman (Kurtman 1961a, b, 1973) assigned Oligocene and Miocene ages to the two different sequences of gypsum (see below). For Cater et al. (1991), the gypsum was late Miocene, correlative with the Eastern Mediterranean Messinian evaporites. The gypsum in the south of the Sivas Basin was Oligocene for Artan and Sestini (1971). The controversies concerning the age of the gypsum deposits were initiated by Kurtman (1961b) himself, when he interpreted the resedimented gypsum around Çorağın Bayırı (near Tepeönü and Fadlun dere), as primary gypsum (Fadlun Dere Formation of Poisson et al. 1996). He discovered lower Miocene fossils in the marls interbedded with the conglomeratic layers and consequently attributed a Miocene age to the gypsum in this site and a general Oligocene and Miocene age to the gypsum at the scale of the Sivas Basin as a whole (with the Hafik gypsum being Oligocene).

Important data were later obtained in the western part of the basin (Gemerek-Ortaköy area), where the base of the red clastics which surround the massive gypsum was dated as Oligocene (mid- to late Oligocene based on vertebrate fauna) (Sümengen et al. 1990). Elsewhere, in the Sivas and Hafik areas, the "massive gypsum" was known to be generally covered, unconformably by transgressive early Miocene marine sediments (Temiz 1994; Poisson et al. 1996). As noted above, the transgressive layers are Aquitanian-aged in the Sivas area. As a result, although the gypsum is considered to be diachronous by some authors, it is well established that the Hafik Formation, the main gypsum sequence in the Sivas Basin, is Oligocene at the top. Even if it is accepted that the Hafik gypsum is well dated, this is not the case for the Fadlun Dere-type deposits (conglomerates and gypsarenites), the uppermost part of which remains undated.

The effects of gypsum remobilisation are rather complicated at the scale of the Sivas Basin. The erosion of the diapirs produced pebbles and sands but also to oversaturated waters which favoured the development of sabkha-like deposits in small basins in a laterally equivalent position to the Fadlun conglomerates. Gypsum also occurs as cement in Quaternary slope breccias. The remobilisation of gypsum was thus an enduring process.

Palaeogeographic evolution of the Sivas Basin

Figures 11 and 12a–h represent an attempt to reconstruct the Cenozoic palaeogeographic evolution of the Sivas Basin from the Palaeocene to the beginning of the late Miocene (before the compressional phase). Several stages are illustrated showing the successive tectonic events that influenced the geometry of the basin and its infill.

Latest Cretaceous–Palaeocene–Early Eocene (Figs. 11, 12a): creation of a fault-controlled deep-marine basin

The Sivas Basin was, by then, clearly individualised as a fault-controlled deep-marine basin, resembling a graben in the midst of shallow-marine platforms, with the Taurus carbonate platform to the south and fluvio-deltaic clastics and nummulitic carbonates to the north (Bahçecik, Karaçayır, Tokuş, North-Gemerek areas). In the basin, the sediments were turbidites (marls, sandstones, conglomerates). Nannofossil assemblages give a Palaeocene–early Eocene age for the bottom of the basin near Ulaş (unpublished data). Extension-related volcanic activity was well developed in the western part of the basin related to deep-seated faults such as the Deliler fault, which probably acted as a normal fault during this period. An olistostrome including volcanic rocks, late Maastrichtian shallow-marine limestones (with *Orbitoides*), recrystallised limestones and various sandstones formed in the Palaeocene interval. The turbiditic flysch and olistostromes emphasise the long-lived tectonic activity along the southern margin of the Sivas Basin.

The late Cretaceous is known in the area of Bünyan in the SW of the basin, where Dirik et al. (1999) described the Tuzla Formation, an olistostromal formation of late Cretaceous to (?) Palaeocene age. Various blocks (limestones, gabbro, ultramafic, metabasalts olistoliths) occur in a matrix of pelagic limestones (calciturbidites), turbiditic sandstone–siltstone and volcaniclastic rocks. The volcaniclastics are associated with andesitic pillow lavas. In the Şarkışla area similar volcanics rocks, dated as Palaeocene, have been interpreted to indicate within-plate volcanism on continental crust (Gökten and Floyd 1987).

Middle and late Eocene (Fig. 12b) (persistence of regional extension; deepening and enlargement of the basin; beginning of emersion)

The extensional tectonics which prevailed during Palaeocene time remained active up to the late Eocene with subsequent volcanism, for instance, in the area of Ortaköy (Fig. 4). Intra-plate volcanism was active throughout the basin, probably related to the activity of the Deliler Fault system. To the south of Ortaköy, mid-Eocene deposits are transgressive on the basement and contain reworked Palaeocene and early Eocene planktic foraminifera. An olistostrome is interbedded in the mid-Eocene layers. This includes late Cretaceous planktic limestones (with *Globotruncana*). This implies that the basin was enlarged and was probably deepening by that time under the control of normal faults. As a result, the basin remained relatively deep and was infilled with turbiditic sandstones and marls with reworked benthic (nummulites) and planktic foraminifera (Bozbel flysch). At the end of Eocene times, the basin became emergent due to a probable regional compressional event.

Early Oligocene (Fig. 12c) (end of extension; regional surrection; evaporitic basin)

This period corresponds to a very rapid and drastic change in the palaeogeography of the basin. After the deep open-marine Eocene flysch deposition, the evaporites of the Hafik Formation were deposited throughout the basin probably in a sabkha-like environment. An early Oligocene age is inferred from the position of the evaporites above the late Eocene flysch and below the mid-Oligocene red clastics (dated by mammals).

Middle Oligocene (Fig. 12d) (reactivation of the basin margins; erosion and sedimentation of red clastics deposits)

Above the evaporites, fluvio-deltaic sedimentation (partly lacustrine) entered the topographically lowest areas of the basin. The sedimentation reflects the persistence of regional tectonic activity, such as the uplift and erosion of the margins of the basin. Salt diapirs underwent growth in an area located to the SW of Sivas. Minibasins were individualised between them and these were the sites of deposition of halokinetic sequences (Ribes et al. 2015).

Late Oligocene (Fig. 12e) (end of erosion; lower rate of sedimentation; enlargement of lacustrine basins between and above salt diapir domes)

The rate of clastic sedimentation diminished, lacustrine areas enlarged, and lacustrine limestones, marls and red clastics were deposited in the minibasins and on top of emergent salt diapirs. Late Oligocene charophytes were discovered in several sites within lacustrine limestones and marls. Some other sites contain mammal fauna, in the Inkonak area, for example, at the southern border of the basin, where they are attributed to the latest Oligocene (MP30 biozone; De Bruijn et al. 1992).

Early Miocene (Fig. 12f) (regional marine transgression; open-marine minibasins, coralgal reef barriers, lagoons)

During the Aquitanian, the central and eastern parts of the basin were transgressed by a shallow sea from the east and southeast. Large areas outside the basin were also flooded, mostly to the north (north of Zara). The western parts of the basin and the southern border, near Kangal, remained emergent or were the sites of lacustrine deposition with coal deposits. The previous morphology persisted, with lacustrine minibasins and uplifted highs. Planktic-bearing marls covered large areas, while coralgal reefs developed as barrier reefs on top of emergent areas, which could be represented by salt diapir domes (e.g. Çaygören). Subsiding, restricted-marine minibasins developed between the barrier reefs, with specific fauna.

Middle Miocene (Fig. 12g) (renewed extension and general emersion; erosion of topographic reliefs: red clastic deposits; lacustrine minibasins with lignites)

Shallow-marine environments persisted until the Langhian. The final retreat of the sea probably occurred during the mid-Miocene. Volcanism was active along the northern and southern borders of the basin, probably related to a new extensional event. K/Ar dating of alkaline basalts gave ages of 10–15 Ma (Parlak et al. 2001). In the Savcun minibasin, the sedimentary sequence includes lacustrine limestones and lignites which have yielded pollen assemblages of mid-Miocene age (Atalay 1993, 1999).

Beginning of late Miocene (Fig. 12h)

Late Miocene deposits are rarely dated. They are known in the Gemerek area as lacustrine limestones and marls, dated by mammals, and to the north of Hafik as lacustrine marls with lignites, also containing a mammal fauna. Near Sivas, the Incesu Formation, dated as late Miocene by mammals, comprises lenses of lacustrine limestones within coarse fluvio-deltaic clastics. These deposits grade upwards into the lacustrine limestones of the Merakom Formation, attributed to the Pliocene. In the Kangal Basin, late Miocene and Pliocene intervals have been dated in the lacustrine sequence which includes important lignitic beds.

A general compressional event occurred soon afterwards resulting in northwest-directed thrusting (i.e. the Sivas thrust and slicing of the basin).

Later (Late Pliocene? Quaternary?)

The initiation of the Kızılırmak graben resulted from a new extensional phase. The renewed erosion fed the graben with red clastics. More recently, the terraces of the Kızılırmak River and tributaries were faulted (mainly SW-NE normal faults) indicating that extensional tectonics persisted.

Connections with the other Central Anatolian Basins and other Middle East Basins

To the west, the connection with the Tuz Gölü Basin is obscured by the Cappadocia Plio-Quaternary volcanics. Dirik et al. (1999) studied the Kayseri area, near Bünyan, which could represent the western extremity of the Sivas Basin. In this area, the pre-Miocene sequences cover the margin of the CACC to the north and the Bünyan metamorphics to the south. Near Bünyan, the late Cretaceous is represented by a deep-marine clastic and volcaniclastic sequence, surrounded by Eocene shallow-marine deposits. The late Cretaceous Tuzla Formation, resting on top of the ophiolites, consists of planktic-bearing turbiditic sandstones, marls and calcarenites. It includes an olistostromal conglomerate with blocks of radiolarites, serpentinites, diabases, various limestones and pebbles of blueschists. Volcaniclastic rocks and basic volcanic rock intercalations are interpreted to reflect a continental within-plate eruptive setting, following the model of Gökten and Floyd (1987) for the Şarkışla area further to the east, within the Sivas Basin. Such magmatism also affects the basement. The Palaeocene is not clearly recognised. The Eocene consists of shallow-marine clastics (conglomerates, sandstones, siltstones) with large benthic foraminifera (e.g. nummulites, alveolines, orthophragmines). As a result, the Bünyan Basin developed related to the same extensional episode as was active in the Sivas Basin. However, the extension was shortlived (latest Cretaceous-(?) Palaeocene) and was probably interrupted by compression by late Palaeocene–early Eocene time. Consequently, the Bünyan Basin did not became very deep or wide (a narrow corridor was suggested by Dirik et al. 1999). However, this remains difficult to determine owing to the shortening (including thrusting) which occured during late Paleogene–early Miocene times. Late Miocene–Pliocene continental deposits (including gypsum and basaltic lava flows) unconformably cover previous structures. The marine lower Miocene transgression of the Sivas Basin did not reach this area. The Kayseri-Bünyan corridor can be effectively considered as the westward termination of the Sivas Basin as it shares many features in common with it. However, after late Cretaceous times, it cannot be interpreted as a deep oceanic basin which could correspond to the Inner Tauride ocean. Further to the SW, the Ulukışla Basin underwent an evolution in some ways similar to that of the Sivas Basin, especially during late Cretaceous–early Eocene times (turbiditic and volcanogenic deposits) (Clark and Robertson 2002). However, the evaporitic event is younger (late Eocene).

To the east, the Sivas Basin is connected with the Erzincan-Tercan Basin which shows a similar shallow-marine lower Miocene bioclastic sequence surrounded by poorly dated continental clastics (Temiz et al. 2002). Further east and SE, marine Miocene also exists in the Muş Basin (Akay et al. 1989) and in the Erzurum Basin which is connected with the Iranian basins and with the Southern Turkish Basins (Robertson et al. 2014).

To the south of the Sivas Basin, the Oligo-Miocene formations are lacustrine (Inkonak area), and a seaway did not exist between Ulaş and Kangal. However, a seaway existed probably towards the Malatya Basin via the Divriği area and from Malatya towards the Eastern Mediterranean, during the Miocene.

Comparison with the northern (Kırşehir and Ankara–Erzincan Suture) and southern (Taurides) areas

Just to the north of Sivas, the basement below the Cenozoic transgression is represented by remnants of the ophiolites obducted from the Neotethyan ocean onto the Kırşehir Massif, and by the Kırşehir Massif itself (probably its eastern termination). The Cenozoic cover of this basement consists of the Tokuş Formation (Özden et al. 1998). This formation, which had a large extension, is composed of a basal coarse, polygenic conglomerates derived from the metamorphic basement and the ophiolites, and also includes richly fossiliferous carbonates with large foraminifera (nummulites). In the Bahçecik area, to the north of Hafik, the sequence is dated as late Palaeocene–early Eocene and the nummulitic limestones as early and middle Eocene. To the north of Hafik, near Düzyayla, the Eocene consists of a flysch-like sequence which contains assemblages of planktic foraminifera of the biozones P 8-P12 (late Palaeocene to Lutetian; unpublished data). To the north of Gemerek, the lower sequence of the Sivas Basin onlaps the Kırşehir Massif. This includes richly fossiliferous, shallow-marine marls and limestones. The assemblage of planktic foraminifera indicates ages ranging from the biozones P9 to P14 (late Ypresien, Lutetian and Bartonian). Similar ages have been obtained on the opposite side of the basin near Ortaköy. As a result, the Palaeocene is not recognised on either side of the basin in this area in contrast to the Şarkışla area to the east. The probable reason is that Palaeocene facies are concealed beneath the Eocene deposits which directly cover the Kırşehir Massif to the north and also the Hinzir Dağ to the south. The same scenario is likely in the Bünyan area. The overall facies pattern during this time can be related to enlargement of the basin, which probably involved deepening related to active normal faulting of the basin margins. Extensional conditions were, however, short-lived as the basin became emergent after the Bartonian.

Towards the south, the environments diversified in the Eastern Taurides. The Taurus autochthon from Sarız, to Gürün and west Hekimhan represents part of a continental fragment which rifted away from the Gondwana during late Triassic–early Jurassic times. The Jurassic and Cretaceous (up to the Cenomanian) are represented by shallow-marine carbonate platforms that were generally surrounded by planktic-bearing deep-marine limestones (Aziz et al. 1982; Robertson et al. 2013). Ophiolitic nappes and ophiolitic mélanges, derived from the closure of the Northern Neotethyan ocean, were emplaced variable distances onto the NE margin of the platforms between Cenomanian and Campanian time. The thrust front of the ophiolitic nappes can be observed in the Hekimhan area. The cover of the nappes in this area is composed of ophiolitic conglomerates and of rudist limestones, dated as late Campanian-lower Maastrichtian. The associated microfauna is also Campanian. The same Campanian deposits can be observed on top of the autochthon to the east of Darende (along the road Darende-Hekimhan) below the nummulitic limestones. In the Hekimhan area, the Maastrichtian sequence is composed of conglomerates, sandstones, marls and limestones and includes a rich microfauna and rudists. The sequence grades upward into an evaporitic sequence in which the K/T boundary has been tentatively identified (Yalçın and Bozkaya 1996).

Shallow-marine carbonate platforms covered the Eastern Taurides uniformly during Palaeocene and Eocene time. They remained unbroken up to the mid-Eocene when ophiolite-derived clastics covered the nummulitic platforms. Afterwards the platforms were broken and dissected into thrust sheets, and then imbricated within a system of nappes that was thrust onto the autochthonous part of these platforms. These events occured after the Eocene (during the Oligocene?), but before the Miocene transgression which covers the contents. This deformation was driven by continent–continent collision of the Kirşehir massif with the Tauride platforms (Robertson et al. 2013). During late Cretaceous–Palaeocene and Eocene times the Sivas Basin differed from the Taurus: it was a deep-marine basin while the Taurus was widely covered by nummulitic carbonates until the mid-Eocene (Booth et al. 2012, 2014). The late Eocene tectonic event in the Taurus and the emersion of the Sivas Basin are thus synchronous. As a result, the Eocene evolution of the Sivas Basin reflects, rather well, the regional tectonic evolution of the Eastern Taurus and Central Anatolia. In these areas, marine deposits dominated during the Eocene time. In contrast, after the Oligocene period of emersion and evaporites deposition, the Sivas Basin was transgressed by shallow-marine seas. Only some parts of the Taurides were affected by this transgression. The Sivas Basin ceased to be connected with the Central Anatolian Basins which became emergent, and it instead became part of the system of marine Middle East basins. The resulting new palaeogeography reflects a new tectonic setting. The Central Anatolian platforms became involved in regional collision and the N–S convergence, and the resulting collisional effects migrated towards the SE where the Southern Neotethyan Basin was in its last stage of closure. The Miocene foreland basins in this area deepened in front of the Tauride thrust front, which collided with the Arabian continent in mid- and late Miocene times. The short-lived Miocene basins in Eastern and South-Eastern Anatolia, including the Sivas Basin, were certainly emergent by this time.

Conclusions

Biostratigraphy

In the Central Paratethys Basins, the mollusc bivalve *Oopecten rotundata* is recognised as an index fossil for the Aquitanian. Relatively frequent in the Sivas Basin (Stchepinsky 1939; Erünal-Erentöz 1956 and our findings), it can also be considered as an index fossil for this basin.

It has been established that the assemblages of Oligocene planktic foraminifera previously described in the Sivas marls are reworked into Aquitanian sediments of this formation.

Reworking of Eocene and Oligocene assemblages of nannoplankton has been also observed in Miocene sequences.

Classical markers (e.g. Mediterranean markers) are absent from assemblages of planktic foraminifera. Other markers are tentatively used instead for the late Burdigalian-lower Langhian, such as *G. lentiana*, *G. ottnangiensis*, *C. chipolensis.*

Generally speaking, the faunal composition resembles that of Paratethys more than that of the Mediterranean Sea.

Chronological evolution of the Sivas Basin

- Latest Cretaceous–Palaeocene–late Eocene period: a fault-controlled extensional open-marine synsedimentary basin persisted.

The basin became emergent by the end of Eocene time (after the Bartonian). This event probably coincided with the final closure of the Northern Neotethyan ocean and the uplift of the Taurus platforms.

- Early to middle (?) Oligocene restricted-marine to evaporitic conditions prevailed and salt layers were deposited (Hafik Formation).
- Mid- to late Oligocene; the environments differed from west to east:
- In the western part of the basin (Gemerek-Şarkışla areas), fluvio-lacustrine conditions prevailed.
- In the central part of the basin, to the east and southeast of Sivas, salt diapirs underwent growth and uplift. Fluvio-lacustrine, halokinetic sequences were deposited in intervening, subsiding minibasins.
- Along the southern margin of the Sivas Basin another minibasin (Selimiye) developed along the northern border of the Taurus platforms. This basin was infilled by a thick sequence of clastic deposits, attributed to the Oligocene. The environment of deposition (marine or continental) is critical to an understanding of the clastic material within the minibasins. It is notable that the Selimiye Basin is located between the minibasins and their inferred source of clastics.
- Beginning of the Miocene, a shallow-marine sea covered large continental areas around the pre-existing Sivas Basin, which was emergent at that time.
- The age of the transgression is here shown to be Aquitanian.
- The transgression did not reach the western and southwestern parts of the basin which remained under lacustrine conditions. The transgression came from the east, where similar Miocene sequences are known.
- The northernmost remnants of Miocene marine deposits are located to the north of Zara. A connection towards the north, towards the Eastern Paratethys, remains questionable and needs more study.

The Sivas Basin was certainly emergent during mid-Miocene time (after the Langhian). This period remains poorly known due to rarity of stratigraphic data. In contrast, the late Miocene is better known due to the mammal fauna in the fluviatile and lacustrine deposits.

The regional character of the transgression is emphasised by the variable topography which was submerged around the previous Sivas Basin: (1) to the north: the Northern Neotethys suture with its Paleogene cover and the emergent parts of the Kırşehir Massif; (2) the central part of the previous Sivas Basin, notably the Oligocene minibasins (Emirhan, Çaygören, Tuzhisar, Karayün, Bingöl, Işhanı-Uzuntepe); (3) to the south: the southern margin of the Sivas Basin in the area of Divriği and the Taurus platforms as well in the Malatya area.

The tectonic evolution of the Sivas Basin since the Oligocene was complicated by salt tectonics which were reactivated during the late Miocene regional compressional event: (1) the Oligocene was the period of deposition of the main evaporitic formation: the Hafik Formation; (2) in the central and eastern parts of the basin gypsum layers in the Miocene sequences were incorrectly interpreted as primary gypsum deposits. These evaporites represent conglomerates and gypsarenites related to an Oligocene salt diapir, which were resedimented within Miocene shallow-marine areas. The remobilised salts gave rise to new evaporitic deposits in small oversaturated basins (sabkhas). As a result, the Miocene, mainly secondary gypsum deposits, differs considerably from the primary Oligocene ones (Hafik gypsum).

In the context of the N–S convergence and collision between the Taurus-Kırşehir and the Eurasiatic plates, the Sivas Basin was alternately subsiding or uplifting: (1) it was subsiding after the emplacement of the ophiolitic nappes and the ophiolitic mélanges during pre-Campanian times; (2) it was uplifted during late Eocene times when the plates finally collided.

The short-lived early Miocene marine episode can be related to a tectonic re-arrangement of the Middle East around the Arabian promontory which underwent collision with the Taurus. The mid- to late Miocene uplift of the Eastern Anatolia followed, leading to renewed erosion and infill of, by then, remnant basins such as the Sivas Basin. The Quaternary system of E–W normal faults delimited the Kızılırmak graben in the region.

The eastern part of the Sivas Basin faced the Northern Neotethyan suture. In contrast, its western extremity appears to end between the CACC and the Bünyan metamorphics and it does not correspond to the Inner Tauride suture which could be, preferentially, located in the southern part of the Taurus platforms similar to the model of Booth et al. (2014), which is also consistent with the model of Sengör and Yilmaz (1981).

Acknowledgments The board of the DARIUS Programme is kindly acknowledged for its support of our research proposals during the years 2010, 2011 and 2012. We also thank Prof. J.P.Callot (Université de Pau et des Pays de l'Adour) and our colleagues from the University Cumhuriyet in Sivas for their help during the fieldwork in the Sivas Basin. Profs. C. Göncüoğlu and Ş. Şen and an anonymous reviewer are acknowledged for their comments which helped us to improve the manuscript. We especially thank Prof. A.H.F. Robertson for editing the manuscript.

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