Geology of the Central Sivas Basin (Turkey)

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ABSTRACT
This paper presents a revised geological map at the 1/50,000 scale of the Central Sivas Basin together with a synthetic stratigraphic chart and cross-sections. The map covers an area of approximately 9840 km² within the Eastern Anatolian orogen. The structure of the studied area is dominated by three major tectonic domains: (i) to the south, a north-verging thrust wedge involving Maastrichtian – Eocene sediments deposited onto an ophiolitic basement, (ii) in the center an Oligo-Miocene domain shaped by salt tectonics detached above the thrust wedge, along a late Eocene salt layer, and, (iii) to the north the Pliocene depocenter onlapping onto the Kirkınır basement. The central halokinetic domain exhibits two generations of minibasins (respectively, early Oligocene and late Oligocene to late Miocene), separated by an evaporite canopy. The map includes new stratigraphic correlations for the pre-salt stratigraphy and improve the comprehension of the southern fold-and-thrust-belt.

1. Introduction

The Sivas Basin in central Anatolia (Turkey) is a foreland basin that formed after the obduction of the tethyan ophiolite during late Cretaceous (Güzou, Temiz, Poisson, & Gürsoy, 1996; Poisson et al., 1996; Temiz, Güzou, Poisson, & Tutkun, 1993). During the late Eocene, the basin disconnection from the marine domain allowed the precipitation of a thick salt layer (Gündogan, Önal, & Depçi, 2005; Kurtman, 1973; Pichat, 2017). Recent studies have highlighted the role of this salt in triggering halokinetic deformations and mini-basin development during the Oligo-Miocene (Callot et al., 2014; Kergaravat, Ribes, Callot, & Ringenbach, 2017; Kergaravat et al., 2016; Pichat, Hoareau, Callot, & Ringenbach, 2016; Pichat et al., 2018; Ribes et al., 2015; Ribes et al., 2017; Ribes et al., 2018; Ringenbach et al., 2013). These works on salt tectonics together with a regional structural study (Legeay, 2017) enable now to provide a new geological map that integrates the Oligo-Miocene halokinetic domains in the fold-and-thrust-belt context.

Based on the compilation and homogenization of existing maps (see Methodology), together with new data acquired during fieldwork, we updated the geological map of the central Sivas Basin (Sheet 1 (Main map)). The map covers the central part of the basin and encompasses stratigraphic levels ranging from the lower Mesozoic basement to the Quaternary deposits. It is complemented by additional data (Sheet 2 (Main map)) highlighting the basin architecture and stratigraphic relationships.

2. Methodology

The present map of 9840 km² is firstly based on the compilation of older studies (e.g. Aktimur, Tekirli, & Yurdakul, 1990; Artan & Sestini, 1971; Kurtman, 1973; Poisson et al., 1996) and follows the 1/100,000 scale geological maps of the areas F23 (Yılmaz, Uysal, Ağan, Göç, & Aydın, 1997), F24 (Aktimur, 1988), F25 (Aktimur & Tütüncü, 1988), G23 (Yilmaz, Sumengen, Terlemez, & Bilgic, 1989), G24 (Atabay & Aktimur, 1997), G25 (Yıldızeli et al., 1984). Based on this compilation, the team, including 4 PhD students (Kergaravat, 2016; Legeay, 2017; Pichat, 2017; Ribes, 2015) and their advisors, performed more than 20 months (cumulated) of field acquisition, involving several thousands of dip-data acquisitions, as well as about 33 km of sedimentological logs.

C. Ribes and C. Kergaravat studied the mini-basins core area and updated the stratigraphic chart of the post-salt deposits with a 1200 km² map at the 1/50,000 scales (e.g. Kergaravat et al., 2016; Ribes et al., 2015; Ribes et al., 2017). This later has been completed.
and extended over a surface of 9840 km$^2$ by E. Legeay and A. Pichat during their respective PhDs (Legeay, 2017; Pichat, 2017; Pichat et al., 2016; Pichat et al., 2018). This work was complemented by the updating of the pre-salt stratigraphic chart thanks to the compilation and homogenization of numerous studies dealing with the pre-salt deposits in different localities of the basin (e.g. Cater, Hanna, Ries, & Turner, 1991; Kurtman, 1973; Poisson et al., 1996).

Finally, the northern domain of the basin, mapped with less detail, follows the study of Özden et al. (1998).

2.1. Geological sheets organization

Two Geological Sheets are included in this paper, accompanied by a georeferenced map (geoTIFF format).

The Sheet 1 (A0 size) presents the geological map at the 1/50,000 scale with the associated legend. Below the main geological map displays (i) a basement map without sedimentary units, (ii) a structural map with the major’s tectono-sedimentary units, and associated tectonic domains and (iii) a sketch map highlighting extension of different evaporite levels.

The Sheet 2 (A1 size) presents the architecture and sedimentary units of the basin. It displays a N–S and E–W correlations charts of pre-salt formations across the central part of the map. Furthermore, basin and regional-scale cross-sections modified from Legeay (2017) are presented to highlight the tectonic features of the Sivas basin.

3. Tectonic setting

3.1. Regional geodynamic

The east–west elongated Sivas Basin (~50 km wide by ~200 km long) developed as a foreland at the boundary of several crustal units (Guezou et al., 1996; Yılmaz & Yılmaz, 2006) (Figure 1). The Pontides to the north have been considered as a part of the Southern Eurasian active margin at least early-mid Mesozoic time (Şengör & Yılmaz, 1981; van Hinsbergen et al., 2016), bounded to the south by a northward-dipping subduction zone that consumed Northern Neotethyan oceanic crust (Okay, 2008). The Izmır-Ankara-Erzincan Suture Zone (IAESZ), which separates the Pontides from the Anatolide-Tauride (Figure 1), contains Middle Triassic to Jurassic ophiolitic remnants ophiolites (see review of Robertson, 2002) and is structurally on top of 85–95 Ma supra-subduction (Dilek, Thy, Hacker, & Grundvug, 1999; Maffione, van Hinsbergen, Gelder, Goes, & Morris, 2017; Parlak et al., 2013; Robertson, 2002; Yalıniz, Göncüoğlu, & Oezkan-Altiner, 2000; Yalıniz & Göncüoğlu, 1998), obducted around 75–85 Ma (see review of Poisson et al., 2016). These Late Cretaceous ophiolites overly the Anatolide-Tauride domain made up of continent-derived units rifted from the northern Gondwana since Triassic time (Fritz de Lamotte et al., 2011; Jolivet et al., 2015; Şengör & Yılmaz, 1981). The Anatolides consist of metamorphosed and exhumed massifs (Kırsızır Block, Tavşanlı and Ayfan zone, and Menderes massif; Figure 1) while the Taurides refer to a belt of non-metamorphosed sedimentary rocks (Plunder, Agard, Chopin, & Okay, 2013; Şengör & Yılmaz, 1981; van Hinsbergen et al., 2016). Several authors suggested that the Kırsızır Block used to be separated from the Tauride block by an oceanic basin, termed the ‘Intra-Tauride Ocean’, and that Late Cretaceous ophiolite were rooted between the two blocks along a suture named the Inner-Tauride suture zone (ITSZ) (e.g. Görür, Oktay, Seymen, & Şengör, 1984; Guezou et al., 1996; Şengör & Yılmaz, 1981). Other authors rather interpret the Kırsızır Block as the northern continuity of the Tauride platform (e.g. Boztuğ et al., 2009; Köksal et al., 2013; Poisson et al., 1996). In the light of recent plate kinematic reconstruction (van Hinsbergen et al., 2016), a unique overriding ophiolitic nappe rooted within the IAESZ is more consistent (Görür, van Hinsbergen, Matenco, Corfu, & Cascella, 2016; Gürer & van Hinsbergen, 2018; Maffione et al., 2017; van Hinsbergen et al., 2016). The ITSZ may form the relict of an Intra-Tauride basin, but there is no strong argument for invoking oceanic lithosphere within (Gürer & van Hinsbergen, 2018; van Hinsbergen et al., 2016).

In Eastern Anatolia, the Anatolide-Tauride/Pontide collision is dated as Late Paleocene to Eocene while Arabia/Tauride collision is dated as Middle to Late Miocene (e.g. Gürer & van Hinsbergen, 2018; Robertson, Parlak, & Ustaömer, 2012). Contractual stages of the Sivas basin during Eocene and Miocene times have been interpreted as the result of the Arabia-Tauride collision, in response to (i) incipient down-going north-dipping subduction of the Southern Neotethys and (ii) retro-foreland before Miocene continental collision (e.g. Dhont, Chorowicz, & Luxey, 2006; Legeay, 2017).

3.2. Sivas Basin

On the basis of previous studies, Figures 2 and 3 summarize the configuration of the three main structural domains of the basin along a synthetic N–S cross-section (Guezou et al., 1996; Kergaravat et al., 2016; Temiz, 1996). Each domain is composed of specific tectono-sedimentary units (Unit 1 to 4), which will be further detailed. From south to north, the basin can be described as followed:

3.2.1. Fold-and-thrust-belt (pre-salt basin, Unit 1)

On the southern edge of the basin, the fold-and-thrust-belt involves Maastrichtian – Paleocene carbonate platforms and associated slope deposits. They cover the
ophiolite (Kurtman, 1973), and farther north, lower to middle Eocene marine flysch formations deposited in a foredeep, north of the growing compressional domain (Artan & Sestini, 1971; Cater et al., 1991; Kurtman, 1973). The fold-and-thrust-belt is covered by a regionally extensive marine salt layer deposited during the late Eocene (Kurtman, 1973; Özçelik & Altunsoy, 1996; Pichat et al., 2016; Pichat, 2017; Tekin, 2001). The salt acted as a passive-roof detachment that decoupled the halokinetic domain of the salt-and-thrust-belt from the thrust wedge below (Figure 3).

### 3.2.2. Salt-and-thrust-belt (salt basin, Units 2 and 3)

In the central part of the basin, the salt-and-thrust-belt involves the late Eocene evaporite and the Oligocene to late Miocene continental to shallow-water marine deposits (Figure 3). In this domain, salt walls and diapirs delineated a first generation of mini-basins (Unit 2) filled by early Oligocene continental clastics (Kergaravat et al., 2016; Pichat, 2017) and capped by an extended Oligocene evaporite canopy (Kergaravat et al., 2017; Ribes et al., 2015; Ribes et al., 2017). Secondary mini-basins (Unit 3) developed on this second evaporite layer and were filled with (i) upper Oligocene continental clastics (Ribes et al., 2015; Ribes et al., 2017) grading to (ii) early Miocene marine sediments, and, eventually (iii) middle to late Miocene continental clastics and reworked evaporites (Kergaravat et al., 2016; Kergaravat et al., 2017; Poisson et al., 2016; Ribes et al., 2017; Ribes et al., 2018). The late Eocene evaporites acted as an efficient decollement level that decoupled the mini-basins province from the compressional wedge at depth, which kept growing and propagating during Oligo-Miocene times (Kergaravat et al., 2016). The first generation of mini-basins (lower Oligocene) and the earliest sediments of the secondary mini-basins (middle Oligocene) recorded a salt-controlled stage of mini-basin initiation and downbuilding. The second generation of mini-basins (late Oligocene to late Miocene) developed while being increasingly influenced by the compressive setting, with squeezing of the bordering diapirs and northward mini-basin tilting (Kergaravat et al., 2017). The northern border of the salt-and-thrust-belt is marked by an east–west trending, north verging thrust, namely the Sivas thrust (Guezou et al., 1996; Poisson et al., 1996; Temiz, 1996) rooted in the Miocene canopy front along the central domain, and in mother salt toward the east and west.

### 3.2.3. Kızılrmak Foreland basin (Unit 4)

The Kızılrmak basin corresponds to the northernmost domain, north of the Sivas thrust (Poisson, Temiz, &
Figure 2. (A) New geological map presented in this study (See Sheet 1 for more detailed map) and location of cross-section (Figure 3) and sedimentary logs (Figure 6). (B) Structural sketch map of the studied area, modified and completed from Kergaravat et al. (2016).
Gürsoy, 1992). It consists of late Miocene to Pliocene sediments onlapping the Kırşehir basement and Campanian to Maastrichtian ophiolitic mélangé (Kılıçlı olistostrome) derived from the northern IAESZ (Yılmaz, Uysal, Bedi, Göç, & Aydın, 1995). To the east, the Kızılırmak basin disappears below the Sivas thrust.

4. Stratigraphy

As previously introduced, the sediment succession within the Sivas Basin has been separated into five depositional groups (Figure 4). Sedimentary infill recorded basin migration from south to north (Figures 5 and Figure 6). On the basis of our own observations and published studies cited below, we shortly describe here the main lithostratigraphic units subdivided in five sedimentary groups. Groups I and II are pre-salt sediments and coeval to the structural unit 1. Groups III, IV and V are post-salt sediments and respectively coeval to the structural units 2, 3 and 4.

4.1. Group I: Maastrichtian – Paleocene

The Maastrichtian – Paleocene succession corresponds to a relative quiet tectonic phase of carbonate platform constructions to the south and turbiditic deposits further north. A general correlation chart of the pre-salt sedimentary formations is proposed in Figure 6.

The Teker Formation (Blumenthal, 1938) (50–700 meters-thick) is dated as Maastrichtian to Thanetian (Inan & Inan, 1990; Yalçın & Inan, 1992). It constitutes the extended and highest relief of the southern edge of the Sivas Basin and directly rests on the ophiolitic basement or above local conglomerates reworking the ophiolite (Kurtman, 1973). The formation starts with tens of meters-thick rudist patch-reefs grading upward to marls and massive dark-grey and fossiliferous carbonates. The formation was interpreted as characterizing a shallow-water carbonate platform (Inan & Inan, 1990; Yalçın & Inan, 1992). Based on the microfauna content, the Paleocene Gurlevik formation described to the east in earlier studies (Kurtman, 1973) is considered as a lateral equivalent to the Teker Formation (Inan & Inan, 1990).

The Yağmurleşmeki Formation (Meshur & Aziz, 1980) (50–300 meters-thick) was formerly attributed to the Maastrichtian to Paleocene, and considered as a lateral equivalent of the Teker Formation (Kavak, Poisson, & Guezou, 1997), but without evidence for lateral facies variations. On the basis of microfauna content (especially *Assilina sp*), we rather suggest a Paleocene – early Eocene age. The formation crops out on the northern side of the Teker Mountain. It comprises a lower basal reddish conglomerates, covered by red clastics which are capped by a ten meter thick carbonate bed (Kavak et al., 1997).

North of the Teker ridge, three other Maastrichtian to late Paleocene formations accumulated in an open deep marine domain according to their facies and microfossil content (Aktimur et al., 1990; Gökten, 1983). These formations include the Kaleköy, Konakyağızı, Çerçekindere and Gazibey formations (Gökten, 1983).

The Kaleköy Formation (Gökten, 1983) (1000–1500 meters-thick) was dated as late Cretaceous to early Paleocene and consists of coarsening upward sandstones rich in volcanic grains, ophiolitic clasts and interbedded with tuffaceous deposits (Gökten, 1983; Gökten, 1986). It outcrops to the western part of the Sivas Basin and at depth in the map area. The formation was interpreted as turbiditic deposits linked to a northward prograding fan system.

The Konakyağızı Formation (Gökten, 1983) (1000–1500 meters-thick) was dated as upper Paleocene (Gökten, 1986) and is exposed in the southwestern part of the basin. It consists of thickening and coarsening upward carbonate-rich debris-flows and sandstones, locally intruded by contemporaneous basalts (Gökten, 1983), and interpreted as turbiditic lobes linked to a northward prograding fan system.
The Çerpaçindere Formation (Aktimur & Tütüncü, 1988) (600 meters-thick) was dated as Maastrichtian to Paleocene (Aktimur et al., 1990) and crops out in the southeastern part of the basin. It includes coarse to fine siliciclastic- to carbonate-rich sandstones locally cut by igneous intrusions and interpreted as turbiditic deposits (Aktimur, Tekirli, & Yurdakul, 1990; Aktimur & Tütüncü, 1988). The formation acts as the eastward lateral equivalent of the Kaleköy and Konakyazi formations.

The Gazibey Formation (Gökten, 1983) (20–50 meters-thick), was dated as late Paleocene, and regionally covers the Konakyazi and Çerpaçindere formations. It is made of reddish shales containing radiolarian fauna and Globorotalia sp. and characterizing an open marine setting.

### 4.2. Group II: Eocene

The Eocene deposits consist of marine turbidites coeval with the initiation of contraction within the Sivas foreland basin. The open-marine basin was confined during the late Eocene, and was filled by evaporites.

The Bahçeçik Formation (Kurtman, 1973) (50–500 meters-thick) was dated as early Eocene to middle Eocene (Kurtman, 1973) (Poisson et al., 1996). Along
the northern edge of the basin, the formation rests unconformably on the IASEZ ophiolitic mélange. Along the southern edge, it is unconformable on the ITSZ peridotites and older sediments. The formation consists of thick debris flow and conglomerates that were fed by the dismantlement of (i) peridotites and carbonates of the Tecer Formation to the south and (ii) Kırşehir metamorphosed rocks and ophiolitic mélange to the north. North of the Gürlevik anticline (near the Aktaş village), the upper boundary of the formation exhibits fining up conglomerates and sandstones marking the transition with the overlying Kozluca Formation (Kurtman, 1973). Along the northern edge of the basin, the Bahçeçik Formation ends with alternating sandstones and marls related to a deltaic environment (Poisson et al., 1996).

We also include in the Bahçeçik Formation the tens of meter-thick Eocene Söğütlü Formation which have only been locally described along the southern side of the Tecer and Gürlevik mountains (Aktımur et al., 1990; Aktımur & Tütüncü, 1988). This also includes debris flows and conglomerates made up of ophiolitic clasts and Eocene carbonate fossils (giving a distinctive cream to grey color) and related to alluvial deposits.

The Kozluca Formation (Kurtman, 1973) (700 meters-thick) was dated early Eocene (Kurtman, 1973). It is exposed in large detachment folds in the southern part of the basin. The formation consists of greenish conglomerates and sandstones made of volcanoclastics and ophiolitic material, alternating with marls, and displaying coarsening-up and thickening-up sequences. The formation was interpreted as turbiditic channel and lobe deposits that were fed from the eastern side of the basin (Artan & Sestini, 1971). The formation acts as a lateral and distal equivalent of the Bahçeçik Formation.

The Tokuş Formation (150 meters-thick) was dated as middle to late Eocene, and only crops out along the northern edge of the basin where it locally developed around Eocene volcanoes (Poisson et al., 1996; Yılmaz et al., 1995). The formation includes sandstones interbedded with nummulitic limestone interpreted as shallow-water marine deposits (Yılmaz et al., 1995).

The Yapalı Formation (Yilmaz et al., 1989) (up to 150 meters-thick) was dated middle Lutetian. It mainly consists of calcarenites and thin-bedded calcareous mudstones and sandstones. Several competent pelagic to conglomeratic limestones beds characterize the top of the formation. These deposits were interpreted as calci- and siliciclastic turbidites.

The Bozbel Formation (up to 700 meters-thick) was dated Lutetian – Bartonian (Kurtman, 1973). It corresponds to regular alternation of thin-bedded sandstones and marls (Artan & Sestini, 1971; Kurtman, 1973). Olistostromes and slumped levels originating from the southern fold-and-thrust-belt are locally present (Figure 6), especially in the western part of the basin where olistoliths can be up to 50 meters-thick. The upper part of the formation become siliciclastic-free with azoic marls and mudstones. The formation was interpreted as characterizing proximal to distal turbidites becoming sediment-starved in the upper part.

The Tuzhisar Formation (no constraint on thickness) was dated late Eocene (Gündoğan et al., 2005; Pichat, 2017; Pichat et al., 2018; Poisson et al., 1996). When preserved from halokinetic deformation, the formation conformably lies over the Bozbel Formation and consists of secondary gypsum beds, locally siliciclastic-rich, displaying tractive sedimentary structures (current ripples, clast gradation, scours …) and interlayered with marls. These beds are capped by a chaotic
mass of porphyroblastic gypsum. The gypsum beds were interpreted as gypsum turbidites and the capping crystalline gypsum as a caprock resulting from the dissolution of former halite deposits (Pichat, 2017). South of the Tecer Mountain, massive gypsum facies also outcrop directly over ophiolite and are interpreted as former shallow-water evaporites precipitated in piggyback basins. The deposition of the Tuzhisar Formation is interpreted as resulting from the tectonic isolation of the basin from the oceanic domain, in an arid to semi-arid climate (Cater et al., 1991; Gündogan et al., 2005; Kurtman, 1973).

4.3. Group III: Early Oligocene

The Selimiye Formation (< 2000 meters thick) is dated as early Oligocene (Güngördan et al., 2005; Kurtman, 1973). Halokinetic deformations induced large thickness variations in the formation (Kergaravat, 2016; Ribes, 2015). It is composed of red shales and
gypsarenites in the lower part, grading upward to alternating red-purple, sandstones and shales. The earliest deposits may have been deposited under a marine shallow-water setting (Gökçen & Kelling, 1985), whereas most of the formation accumulated in a continental playa-lake environment (Gökçen & Kelling, 1985; Ribes, 2015).

4.4. Group IV: Middle Oligocene to Middle Miocene

The Group IV includes the Karayün, Karaçaoren and Benilkaya formations (e.g. Ribes et al., 2017; Ribes, 2015). Due to halokinesis, these successive formations are sometimes found incomplete (Kergaravat et al., 2017). Moreover, their facies repartition in the mini-basins is controlled by salt flow as well as by the increasing activity of south-verging thrusts (Kergaravat et al., 2016; Ribes, 2015; Ribes et al., 2018).

The Karayün Formation (< 2500 meters thick) was deposited between middle Oligocene and late Oligocene. Currently, only the top part of the Karayün formation is dated as late Oligocene by biostratigraphy (Poisson et al., 1996). This formation is always cropping out above the Oligocene salt canopy, with the exception of the western end of the Sivas basin (Tatlıçak area), where the Karayün basal contact is conformable on the Selimiye Formation (Yılmaz, Uysal, Ağan, & Göç, 1997). In the mini-basins, the formation involves three successive members (Ribes et al., 2017): (i) a lower member made of mudstone interbedded with sandstones and evaporite beds; (ii) a middle member made of amalgamated channelized sandstones and conglomerates; and (iii) an upper member made of mudstone with few isolated sandstones, interbedded with lacustrine carbonate and evaporite beds. These deposits characterize a distributary fluvial system with each member, respectively corresponding to (i) playa lake and distal terminal splay deposits, (ii) fluvial braided deposits and (iii) saline lacustrine deposits (Ribes et al., 2017).

The Karaçaoren Formation (< 2500 meters thick) was dated as Aquitanian to Burdigalian (Ribes, 2015; Sirel, Ozgen Erdem, & Kangal, 2013). Along salt ridges, it unconformably overlies the Karayün Formation, the evaporite canopy, and, farther east, the Selimiye Formation (Kergaravat et al., 2016). Away from evaporite bodies, the contact is generally concordant over the Karayün Formation or discordant above the Selimiye Formation. North of Sivas thrust, the Karaçaoren Formation is deposited over deformed Eocene deposits (Guezou et al., 1996; Poisson et al., 1996). The Karaçaoren formation is made of bioclastic sandstones, grainstones, recifal patches, marls and evaporites. It records a marine transgression with facies deposited in a mixed deltaic and carbonate ramp environment evolving to restricted coastal bays (Ribes, 2015; Ribes et al., 2018).

The Benilkaya Formation (< 1000 meters thick) is dated from late Burdigalian to Tortonian (Poisson et al., 2016; Ribes, 2015). It is mostly observed in the central part of the map, around the most uplifted and well-exposed secondary mini-basins filled by the Karayün and Karaçaoren formations. The formation is conformable over the Karacaören Formation, or unconformable over older formations along the limit of the halokinetic domain. The lower part of the formation is composed of a fining-upward coarse conglomerates and sandstones related to a distributive fluvial system (Ribes et al., 2018). The upper part of the formation displays mudstones, carbonates beds and evaporites deposited in saline mudflat to saline lacustrine environment (Ribes et al., 2018).

4.5. Group V: Late Miocene – Pliocene

The upper part of the sedimentological column is made of upper Miocene and Pliocene continental deposits corresponding to the present-day foreland domain, and including the Incesu and the Merakom formations. These are well represented to the North of the Sivas Basin, forming the Kızlırmak basin, where the sedimentary thickness does not exceed a few hundred of meters (Guezou et al., 1996). To the south, the time equivalent Kangal Formation also covers older formations, forming cuesta-like topography in the Kangal Basin with south directed, low dipping to tabular stratifications.

The lower boundary of the Incesu Formation is dated as upper Miocene (Poisson et al., 1996). It consists of sandstones and conglomerates, with local intercalations of white marls and lacustrine limestones deposited in an alluvial environment.

The Merakom Formation is dated as lower Pliocene (Poisson et al., 1996) and is unconformably deposited over the Incesu Formation. It is essentially made of lacustrine facies with limestone beds and green marl intercalations.

5. Conclusion

The geological map and the stratigraphic chart displayed in this paper act as a major regional synthesis of the Central Sivas Basin.

If the post-salt and halokinetic history of the basin had recently been updated, the pre-salt stratigraphic setting was up-to-now remaining quite confusing regarding the literature, with many formations that had been independently defined by different authors in different localities. The homogenization of the pre-salt stratigraphic units and their correlations enable to better understand 40 Ma of the basin history (corresponding to ~5 km of sediments). It emphasis the lateral facies and thickness variations of the different formations and highlights a deep...
marine foredeep, affected by volcanic activity, and extensively filled by siliciclastic, carbonate-rich and volcanoclastic turbidites. These later were mainly fed from (i) the carbonate platforms and basement rocks dismantling in the growing southern fold-and-thrust-belt initiated at Early Eocene, and (ii) north deltaic input along Kirşehir block and Pontides culminations.

The new geological map, together with the complementary documents improves thus the comprehension of the tectono-sedimentary evolution of the Sivas Basin, even if the stratigraphic chart remains unchanged for the post-salt formations. Our extended geological mapping of the central halokinetic domain enables to appreciate a large panel of new halokinetic structures (welds, diapirs, salt sheets and canopies) and salt-walled mini-basins. Their full description remains beyond the scope of this study but they should soon encourage further sedimentary and structural investigations of the Oligo-Miocene halokinesis in the Sivas Basin.

Software

Outcropping geological surfaces were mapped in the field with the support of orthophotographs provided by GoogleEarth®. The team used compass-clinometers or Fieldmove™ (Middland Valley) on a tablet to collect dip data. The field data were geo-referenced and integrated into a digital environment (Quantum GIS) where the geological limits were extrapolated from satellite images. Final editing of the Geological Map was made using Adobe Illustrator and MAPublisher. The proposed cross-section were constructed and restored using 2DMove® (Middland Valley).

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