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# Factors controlling stratal pattern and facies distribution of fluvio-lacustrine sedimentation in the Sivas mini-basins, Oligocene (Turkey)

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## ABSTRACT

The Sivas Basin, located in the Central Anatolian Plateau of Turkey, is a foreland basin that records a complex interaction between sedimentation, salt tectonics and regional shortening during the Oligo-Miocene leading to the formation of numerous mini-basins. The Oligocene sedimentary infill of the mini-basins consists of a thick continental succession, the Karayün Formation, comprising a vertical succession of three main sub-environments: (i) playa-lake, (ii) fluvial braided, and (iii) saline lacustrine. These sub-environments are seen as forming a large Distributive Fluvial System (DFS) modified through time as a function of sediment supply and accommodation related to regional changes in climate and tectonic regime. Within neighbouring mini-basins and despite a similar vertical stratigraphic succession, subtle variations in facies assemblages and thickness are observed in stratigraphic units of equivalent age, thus demonstrating the local control exerted by halokinesis. Stratigraphic and stratal patterns reveal in great detail the complex interaction between salt tectonics and sedimentation including different types of halokinetic structures such as hooks, wedges and halokinetic folds. The regional variations of accommodation/sediment supply led to coeval changes in the architectural patterns recorded in the mini-basins. The type of accommodation regime produces several changes in the sedimentary record: (i) a regime dominated by regional accommodation limits the impact of halokinesis, which is recorded as very small variations in stratigraphic thickness and facies distribution within and between mini-basins; (ii) a regime dominated by localized salt-induced accommodation linked to the subsidence of each individual mini-basin enhances the facies heterogeneity within the DFS, causing sharp changes in stratigraphic thickness and facies assemblages within and between mini-basins.

## INTRODUCTION

The distribution, type and stacking pattern of fluvio-lacustrine units within a Distributive Fluvial System (DFS) in an endorheic basin are controlled by several autocyclic and allocyclic processes (Huerta *et al.*, 2011; Weissmann *et al.*, 2013). A DFS, generally forms at margins of sedimentary basins where water that enters into the basin through a feeder channel or valley becomes unconfined and free to avulse or/and bifurcate, forming radial channel pattern (Hartley *et al.*, 2010; Weissmann

*et al.*, 2010). However, DFS do not always have a fan-shape (Hartley *et al.*, 2010). There are close similarities between modern (Hartley *et al.*, 2010; Weissmann *et al.*, 2010, 2013; Davidson *et al.*, 2013) and ancient examples of DFS (Friend, 1977; Parkash *et al.*, 1983; Stear, 1985; Kelly & Olsen, 1993; Nichols & Fisher, 2007; Cain & Mountney, 2009). A DFS is sensitive to a combination of factors more or less directly related to the palaeoclimatic, palaeogeographic and tectonic setting (i.e. upstream controls; Hartley *et al.*, 2010). The complex relationships between subsidence, sediment supply, base level and climatic change considerably modify the sedimentary record observed in a DFS (Weissmann *et al.*, 2013). Climate changes can affect vegetation and rainfall, which in turn affect erosion, sediment source, and discharge not only in the sedimentary basin but also in the sediment source-areas, influencing the rate and the pattern of sediment

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supply (Shanley & McCabe, 1994; Huerta *et al.*, 2011). Tectonic events in the source area can also modify the sediment supply, and tectonically driven regional subsidence governs the overall accommodation (e.g. Catuaneu *et al.*, 2011).

The development of a DFS in a salt province will record a complex interplay between the classical controls of sedimentation in a basin and the local influence of salt tectonics. The sedimentary successions deposited on top of thick salt layers commonly result in the generation of mini-basins in response to differential subsidence related to salt flow initiated and driven by sediment loading and tectonics (Hudec & Jackson, 2007; Hudec *et al.*, 2009; Trudgill, 2011). The initiation and evolution of salt-related mini-basins involves a variety of local and regional controls. Local factors include salt withdrawal, which influences the rate and style of subsidence, and salt-structure growth, which influences the topographic expression of growing diapirs (Andrie *et al.*, 2012; Giles & Rowan, 2012). The growth of salt diapirs at the surface, which reflects a subtle interplay between sedimentary loading and local stresses, exerts a control on the dispersal and distribution of sediment across the mini-basins (Matthews *et al.*, 2007; Trudgill & Paz, 2009; Trudgill, 2011; Banham & Mountney, 2013a,b; Venus *et al.*, 2015). Thus, mini-basins will record the interplay between the history of the overall sedimentary system and the sedimentary response to contemporaneous local salt movement (Hodgson *et al.*, 1992; Barde *et al.*, 2002; Prochnow *et al.*, 2006; Buck *et al.*, 2010; Newell *et al.*, 2012; Banham & Mountney, 2013c).

Relatively few continental mini-basins are known and few outcrop examples have been described in the literature. Rare subsurface examples have been studied using 2D/3D seismic data sets, such as the Triassic Skagerrak Formation in the UK sector of the Central North Sea (Hodgson *et al.*, 1992; Smith *et al.*, 1993; McKie *et al.*, 2010), the Precaspian Basin in Kazakhstan (e.g. Barde *et al.*, 2002), and the Paradox Basin in USA (e.g. Kluth & DuChene, 2009). Few examples are known at outcrop showing the associated fluvio-lacustrine infill. Among such mini-basins, the best published examples are the Eocene Carroza Formation of the La Popa Basin, deposited in an isolated mini-basin (Mexico; Buck *et al.*, 2010; Andrie *et al.*, 2012; Giles & Rowan, 2012), the Pennsylvanian–Jurassic Formations in the Paradox Basin (USA; Bromley, 1991; Lawton & Buck, 2006; Prochnow *et al.*, 2006; Matthews *et al.*, 2007; Trudgill, 2011; Banham & Mountney, 2013b,a; Venus *et al.*, 2015) and the Permian–Cretaceous Precaspian Basin (Kazakhstan and Russia; Barde *et al.*, 2002; Newell *et al.*, 2012).

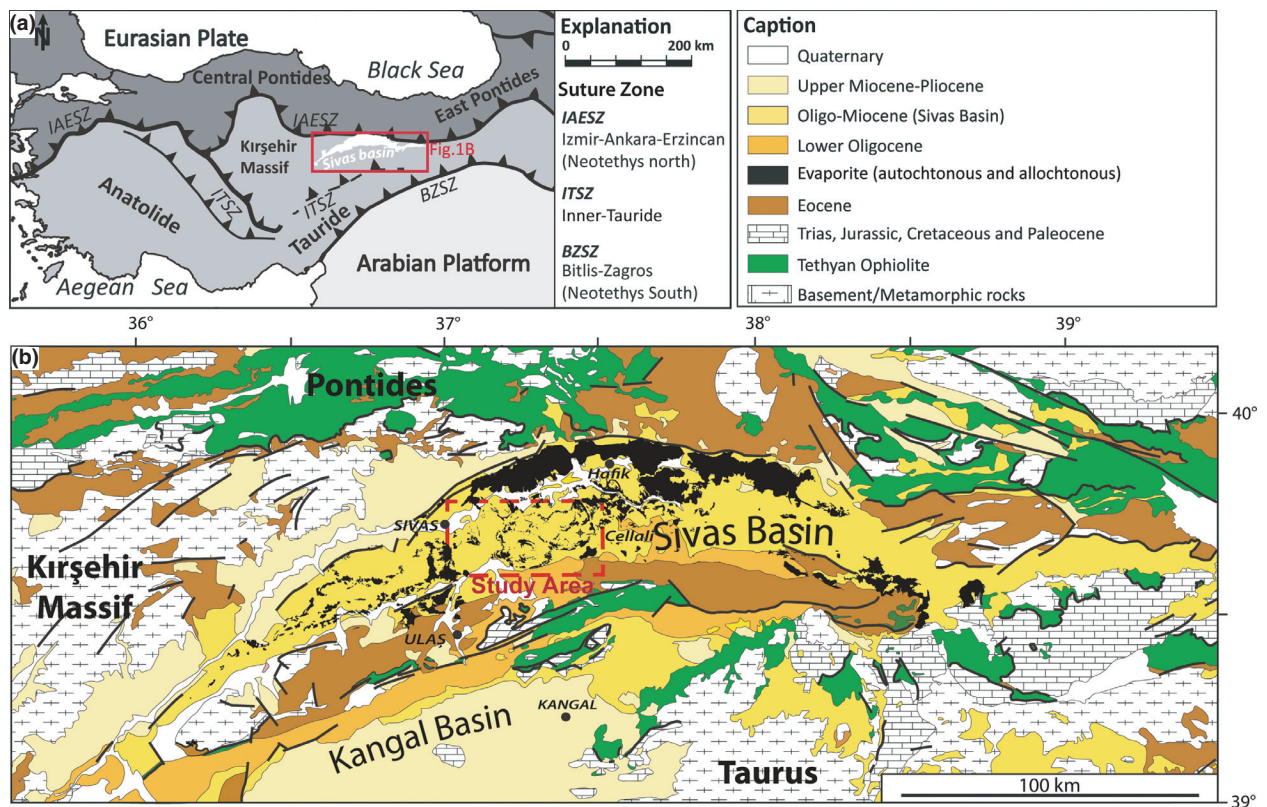
The spectacular assemblage of salt mini-basins described here in the Sivas Basin (Central Anatolia, Turkey; Fig. 1) is mainly Oligo–Miocene in age (Ringebach *et al.*, 2013; Callot *et al.*, 2014). At least twenty mini-basins compartmentalize the central part of the Sivas Basin and expose the entire sedimentary pile largely affected by salt tectonics (Fig. 2). In a previous study

(Ribes *et al.*, 2015), the Emirhan mini-basin was described in great detail using satellite imagery and outcrop mapping. The reference section and the depositional environments of the mini-basin infill, represented by the Oligocene Karayün Formation, was clearly established. Outcrop investigations focussing on the Emirhan mini-basin allow us to distinguish three long-lived vertically stacked depositional systems within a large DFS: (i) playa-lake and distal terminal splay, (ii) fluvial braided system with low sinuosity and, (iii) saline lacustrine deposits. During deposition of the Karayün Formation, halokinesis had an important influence on the geometry of the individual sedimentary sequences and facies distributions within the sequences filling the Emirhan mini-basin. This mini-basin also recorded significant temporal variations in the stratigraphic assemblage attributed to apparent allocyclic factors. However, additional data was acquired to define more precisely the relative importance of these allocyclic controls and improve our regional understanding of the Sivas mini-basins.

This paper provides a comprehensive description and mapping study of ten adjacent mini-basins, with the purpose addressing the interaction between allocyclic regional factors and local salt-induced factors in comparison with the previously described Emirhan mini-basin. The aim of our study is to focus on the relative influence of regional factors and halokinesis on a DFS during the development of salt-related mini-basins, based on spatial and temporal changes in the facies assemblages, stratigraphic architecture and thickness, as well as stratal structures related to salt movement (i.e. halokinetic structures). In this study, we develop a new tectono-stratigraphic–paleoenvironmental model for the evolution of the Sivas mini-basins and provide a better understanding of the influence of regional factors on the DFS developed simultaneously in the different studied mini-basins. Moreover, we attempt to understand how halokinesis isolates neighbouring mini-basins and produces salt-induced local accommodation that can modify the sedimentary record. Finally, we discuss how tectonically driven regional accommodation can be differentiated from salt-induced local accommodation.

## GEOLOGICAL SETTING AND STRATIGRAPHY

The Sivas Basin is located at the junction between three crustal terranes: the Pontides orogenic belt to the north, the Kırşehir metamorphic massif to the west and the Taurides to the south, derived from the Gondwana continent (Fig. 1a; Cater *et al.*, 1991; Yılmaz & Yılmaz, 2006). The Sivas Basin formed after the closure of the north Neotethys, above the obducted ophiolite and the Kırşehir and Taurus blocks (Yılmaz & Yılmaz, 2006; Rolland *et al.*, 2010). The sedimentary record starts during the Late Cretaceous–Paleocene with the development of a carbonate platform along the southern margin of the Sivas



**Fig. 1.** (a) Tectonic setting of Turkey, with main continental blocks, major suture zones and Oligocene–Miocene Sivas basin deposits, modified from Okay *et al.* (2006). (b) Geological map of the Sivas Basin with the three major tectonic units (Pontides, Taurus and Kırşehir massif) and location of the study area, modified from Baykal & Erentöz (1966). Distribution of evaporite formations (in black) in the Sivas basin, showing polygonal pattern of outcrops in the study area (Fig. 2).

Basin directly on the ophiolite (Gürlevik Dağı and Teçer Dağı limestones; Figs 1b and 2a; Kurtman, 1973; Cater *et al.*, 1991). During the Paleocene to Eocene, the Sivas Basin records a general deepening of its southern boundary related to the onset of Tauride shortening with the development of a foreland basin (Kurtman, 1973; Guezou *et al.*, 1996; Poisson *et al.*, 1996; Kavak *et al.*, 1997; Görür *et al.*, 1998; Yılmaz & Yılmaz, 2006; Onal *et al.*, 2008). The lower part of this succession is thought to have been deposited in a quiet marine context in a basin belonging to the obducted ophiolitic basin, while the upper part is a flysch type deposit in a context of north-verging shortening, north of the Taurus.

An olistostrome dated as Early Eocene and the Eocene Bahçecik conglomerate with clasts of serpentinite, peridotite, marble, red chert, limestone and quartz may record the onset of the first compressive event on the southern margin of the Sivas Basin (Fig. 2a; Kurtman, 1973). The distal facies of the Eocene succession consist of deep-marine turbidites and volcanoclastic sediments corresponding to the Bözbel Formation, deposited in an East–West elongated foreland basin (Fig. 2b; Kurtman, 1973; Özçelik & Altunsoy, 1996). The upper part of the Eocene flysch succession comprises a calcareous mudstone succession, passing abruptly upward into a coarser and thicker shallow-marine facies (Fig. 2a; Kurtman, 1973; Cater *et al.*, 1991). This shallowing-up succession

is capped by an evaporitic unit, which is the mother salt deposit (Fig. 2a; Özçelik & Altunsoy, 1996; Gündogan *et al.*, 2005). In a few places where the section appears relatively unaffected by halokinesis, the evaporite unit is composed of alternating beds of anhydrite and gypsum, named the Tuzhisar Formation (Fig. 2a; Özçelik & Altunsoy, 1996; Gündogan *et al.*, 2005). The Tuzhisar Formation is overlain by red-purple to green fine-grained sandstone and shale attributed to a playa-Lake environment and called the Selimiye Formation (Kurtman, 1973; Poisson *et al.*, 1996).

After deposition, salt tectonics resulted in large-scale salt movements and the initial salt stock appears largely remobilized to above the Selimiye Formation, forming allochthonous salt sheets and bodies transported northward below younger mini-basins (C. Kergaravat, J.-P. Callot, C. Ribes & J.C. Ringenbach, submitted). A 1500-m-thick succession of interbedded shale and salt/anhydrite was encountered in the Celali-1 borehole below the Selimiye Formation and is visible in the subsurface on unpublished seismic data, demonstrating the widespread remobilization of the so-called autochthonous Tuzhisar salt Formation (Özçelik & Altunsoy, 1996; Gündogan *et al.*, 2005; Onal *et al.*, 2008). The whole basin structure has been reconsidered based on larger scale field work integrating part of the unpublished seismic data set (C. Kergaravat, J.-P. Callot, C. Ribes & J.C. Ringenbach,

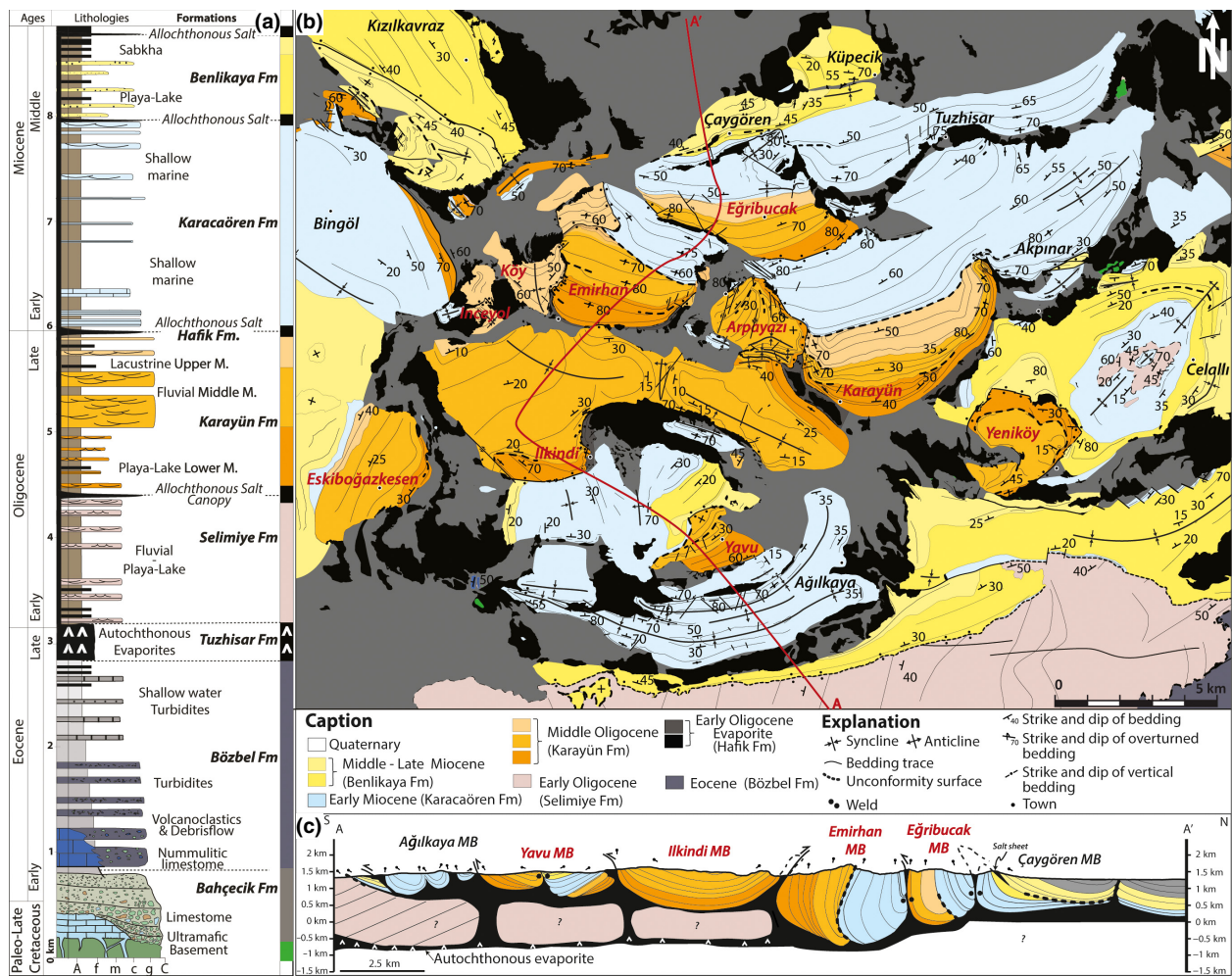


Fig. 2. (a) Regional composite lithostratigraphic column showing average thickness of the various stratigraphic units in the central Sivas Basin. (b) Geological map of the central Sivas Basin, showing location of gypsum and anhydrite outcrops (black and grey) and the polygonal geometry of gypsum walls surrounding the mini-basins. Mini-basins names in red are the studied mini-basins: Emirhan, Karayün, İlkindi, Arpayazi, Eğribucak, Yeniköy, Yavu, Eskiboğazkesen, Köy, and Inceyol. (c) Schematic South-North cross-section across the Sivas mini-basins.

submitted and Legeay *et al.* in prep). The present-day deep basin geometry shows a north-verging triangle zone below the autochthonous salt (remobilized Tuzhisar Formation), which has acted as a décollement. Following the deposition of the Selimiye Formation, the depocentre migrated northwards as the salt became remobilized into a canopy located above the Selimiye Formation and below the Karayün and Karacaören Formations. This canopy is visible on seismic and can be mapped in the eastern part of the Sivas Basin south of Zara (Fig. 1). This canopy is widely exposed in the northern part of the Sivas Basin (Fig. 2b) and is locally overlain by small patches of Karacaören shallow marine deposits. This northern evaporitic unit is called the Hafik Formation (Fig. 2b; Kurtman, 1973). The youngest salt sheet, not yet totally separated from the former occurs above the Benlikaya Fm.

In the central part of the Sivas Basin, the canopy is overlain by a thick succession of red beds of the fluvio-lacustrine Karayün Formation, reaching a maximum thicknesses of 2400 m (Fig. 2b; Kurtman, 1973; Cater

*et al.*, 1991; Poisson *et al.*, 1996; Ribes *et al.*, 2015). The deposition of the Karayün Formation coincided with the onset of salt tectonics and the subsidence of mini-basins lying on top of the basal canopy in the central part of the Sivas Basin (Orange Fig. 2b; Ribes *et al.*, 2015). Above the Karayün Formation, the deposition of the shallow-marine Karacaören Formation marks a regional transgression during the Early Miocene (Kurtman, 1973; Cater *et al.*, 1991; Özcan *et al.*, 2009; Sirel *et al.*, 2013), expressed as a conformable contact in some places (centre of the mini-basins), whereas the marine beds onlap the folded Karayün Formation (in blue, Fig. 2b) at numerous locations (edges of the mini-basins). The Karacaören Formation contains shallow-marine marls and sandstones that are interbedded locally with algal limestone (Fig. 2b; Kurtman, 1973; Özcan *et al.*, 2009; Poisson *et al.*, 2010, 2011; Sirel *et al.*, 2013). The overlying Benlikaya Formation is characterized by alluvial to fluvial deposits of Middle Miocene to Early Pliocene age, composed of reddish sandstone, conglomerate and interbedded mudstone (yellow Fig. 2; Poisson *et al.*, 1996, 2010).

The original evaporite level contains halite, as shown by the presence of salt springs (i.e. the Fadlun Tuz area and Bingöl Lake) and borehole data from the Cellali-1 well (Onal *et al.*, 2008). Near the surface, however, the diapiric material only contains sulfate residuum (i.e. caprock) left over after the emplacement of the salt structure and salt withdrawal.

**DATA AND METHODS**

The paper is mainly based on detailed analyses of outcrop data and geological interpretation of colour composite images from the Geoeeye satellite. More than 12 km of vertical sedimentological sections were measured across the studied mini-basins (Fig. 3). The thickness variations and lithological features of the fluvio-lacustrine section were collected in the field and mapped onto the satellite image. This was done to characterize the lateral and vertical facies relationships as well as the architecture of sandstone bodies, as well as the thickness variations and stratal patterns of the stratigraphic members across each mini-basin and between them. These mini-basins offer the most complete exposure of the stratigraphy and stratal

patterns of the Karayün Formation, due to the varying degrees of tilting such as observed in the Emirhan mini-basin, which is tilted to vertical (Figs 2 and 3).

The description of lithofacies and facies associations, as well as the criteria for the recognition of architectural elements of fluvial origin, follows the classification and interpretation scheme established by Miall (1985).

A total of 633 palaeocurrent measurements were collected on 16 sites across the 10 mini-basins, based on cross-bedding foreset azimuths, flute marks and parting lineation, to determine the paleoflow orientation and potential source area of sediment supply (Fig. 3). The methodology due to Lindholm (1987) was used to calculate the mean vector.

**KARAYÜN FORMATION REGIONAL STRATIGRAPHY**

**Overview of previous studies of the Karayün Formation**

The Karayün Formation was defined in the vicinity of Karayün village (Cater *et al.*, 1991; Poisson *et al.*, 1996), and a brief description of the vertical section in terms of

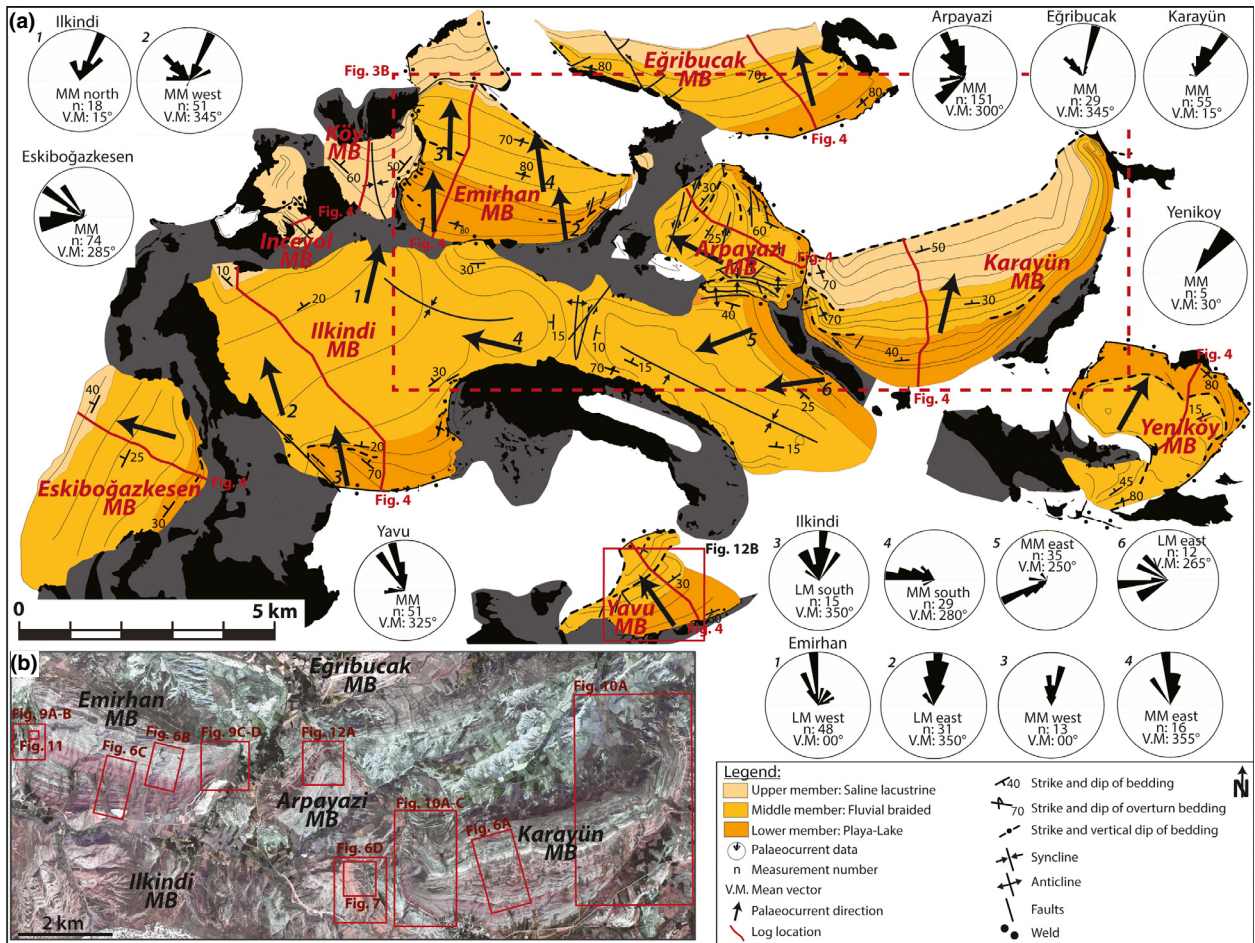


Fig. 3. (a) Geological map of the central Sivas Basin, showing location of measured sedimentological sections and palaeocurrent data. Dashed-line rectangle area covered in (b). (b) Geoeeye satellite colour composite image of some Sivas mini-basins showing location of panoramic views and interpreted zooms given on Figs 8, 9, 11–14.

an accurate interpretation of the depositional environments was given by Çiner et al. (2002). Unfortunately, the Eğribucak section was interpreted erroneously as a younger additional unit. Precise structural mapping along the same section and facies analogies demonstrate the repetition of the succession cropping out in two adjacent mini-basins separated by a salt weld (Ribes et al., 2015). The reference stratigraphic section has been adjusted and modified consequently to take into account this structural complexity.

The regional stratigraphy sub-division of the Karayün Formation adopted here has been described in a previous publication focusing on facies interpretation, as well as the stratal and stacking patterns and regional depofacies observed in the reference Emirhan mini-basin (Ribes et al., 2015).

Three long-lived depositional systems have been recognised from base to top of the Karayün:

- (1) Lower member: playa-lake and distal terminal splay deposits corresponding to Unit 1, 2 and 3 in the Emirhan mini-basin.
- (2) Middle member: fluvial braided deposits corresponding to Unit 4 and 5 in the Emirhan mini-basin.
- (3) Upper member: saline lacustrine deposits corresponding to Unit 6 in the Emirhan mini-basin.

These three depositional systems fill ten mini-basins in the central area and correspond to large-scale genetically related stratal packages that are separated by basin-wide

surfaces expressed by changes in texture and facies assemblages, as well as stratal and stacking patterns (Figs 3 and 4). The facies associations of these three members and the transitions between them are detailed below based on the study of the Lower, Middle and Upper members in ten mini-basins named after local villages, i.e.: Emirhan, Karayün, Arpayazi, Eğribucak, Ilkindi, Eskiboğazkesen, Yavu mini-basin, Yeniköy, Inceyol and Old-Emirhan (Figs 3–5).

### Lower member of the Karayün Formation: red oxidized playa-lake and sheet flood deposits

#### Description

The Lower member is a dominantly reddish shaly interval with interbedded sheet flood and playa lake deposits developed at the base of most of the studied mini-basins (Fig. 3). This Lower member reaches a maximum thickness of 1200 m, with varying proportions of floodplain mudstone (65–80%), containing interbedded sandy fine grained single-storey sheet-like bodies with common conglomerates (20–30%) and evaporite beds (0–5%; Figs 4–6d, 7 and 8c, d). The mudstone facies consists of red-oxidized purple to green laminated (Fl; Fig. 4) or massive mudstones (Fm; Fig. 4) forming continuous horizons that can be correlated. The interbedded red to light-brown fine-to

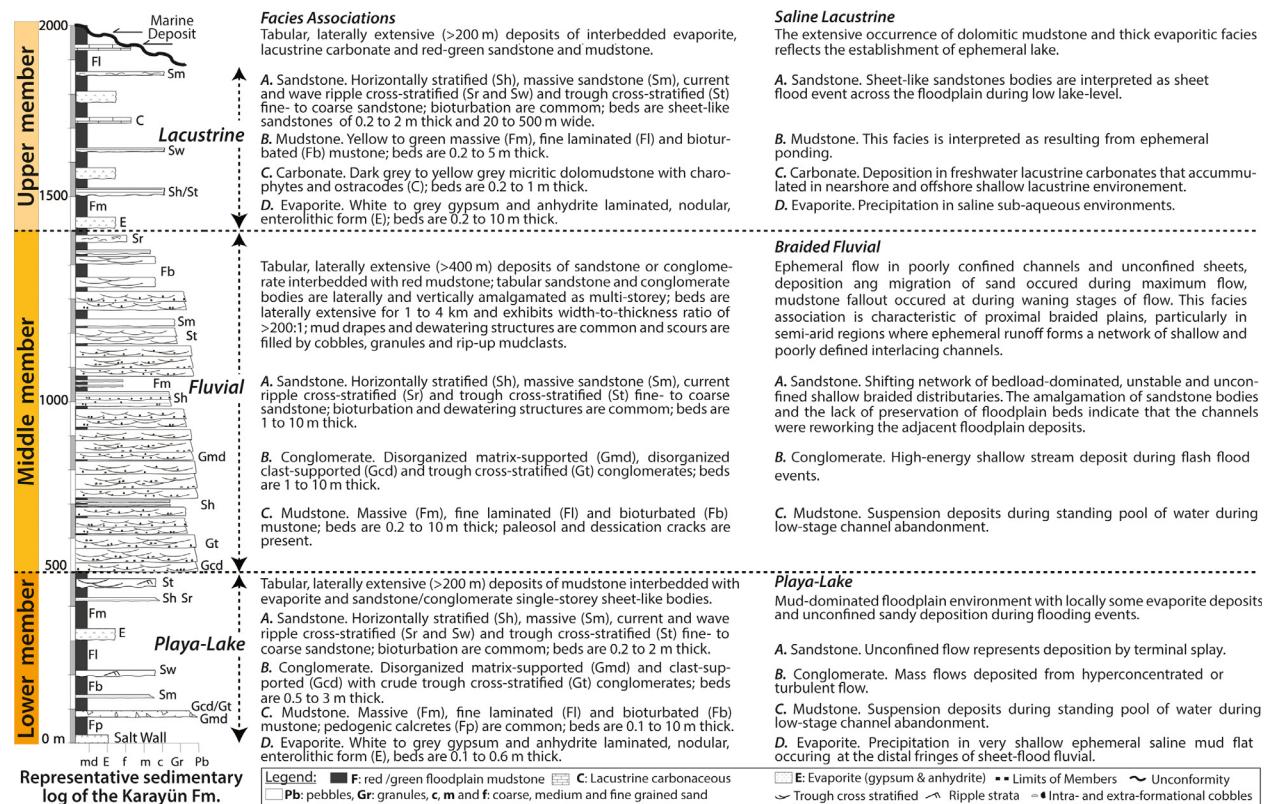


Fig. 4. Description and interpretation of lithofacies and stratal architecture of the Karayün Formation in ten mini-basins, with a representative sedimentary log of sections measured across the Karayün Formation.

coarse-grained sandstone beds make up isolated single-storey bodies with sheet-like geometries and sharp bases with clay chips. Beds range from 0.2 to 2 m thick and typically show a 500 m average lateral extension. Most of these beds display small-scale climbing (Fig. 6d), symmetrical and asymmetrical ripple (Sr) laminations (Sr and Sw; Fig. 4), as well as trough-cross-bedded (St; Fig. 4) and parallel laminations (Sh/SI; Fig. 4). Uncommon sandstone beds appear as massive or structureless (Sm, Fig. 4). These sandstone bodies are equivalent to the laminated sand sheets (LS; e.g. Miall, 1985) and sandy bedforms in the sense of Miall.

Conglomerates occur either as structureless bodies with a sheet-like geometry 20–50 m long and 0.5–1.5 m thick containing matrix-supported clasts (Gmd; Fig. 4), or as conglomeratic beds 0.5–3 m thick containing clast-supported gravel with medium-scale trough cross-stratification in tabular beds showing a lateral extent of 50–800 m. Both types of conglomerate show a basal scoured surface (Ss; Fig. 4) capped by red intraformational mudstone pebbles and conglomerate clasts (GB; Fig. 4). Conglomerate clast types include limestone, serpentine, peridotite, quartzite and evaporite pebbles (1–8 cm in diameter). Anhydrite and gypsum occur as dm-thick laminated or enterolithic layers associated with clay laminations, and form continuous mappable horizons.

Interpretation

The sheet-like geometry of sandstone and conglomerate beds is attributed to the diffuse flow of unchanneled sheet floods across the alluvial plain (e.g. Tooth, 2000, 2005; North & Davidson, 2012). The conglomerate beds are characteristic of mass flow deposits, interpreted as a product of rapidly decelerating high-magnitude flow associated with hyper-concentrated flash-discharge events on a floodplain (e.g. Miall, 1996). This interpretation is supported by the presence of sheet-like structures, gravel clustering, poor sorting and the absence of internal stratification (e.g. Nemec & Steel, 1984; North & Davidson, 2012). Planar-parallel-laminated sandstone beds indicate rapid episodes of sand deposition from high-velocity unidirectional flows and suggest dominant bedload transport (e.g. Hampton & Horton, 2007). Climbing ripples and ripple cross-bedding in the upper fine-grained intervals of individual sandstone beds indicate decreasing discharge during flood events (e.g. Miall, 1985). Also, small-scale Srs in the siltstone–mudstone alternations are thought to be formed by wind-driven waves in water depths of a few cm, probably within small ponds (e.g. Allen, 1968). The vertical facies pattern, showing a succession of red and green mud and evaporite facies, can be attributed to successive progradation and retrogradation of the distal part of the system (e.g. Jackson *et al.*, 1990), or might be con-

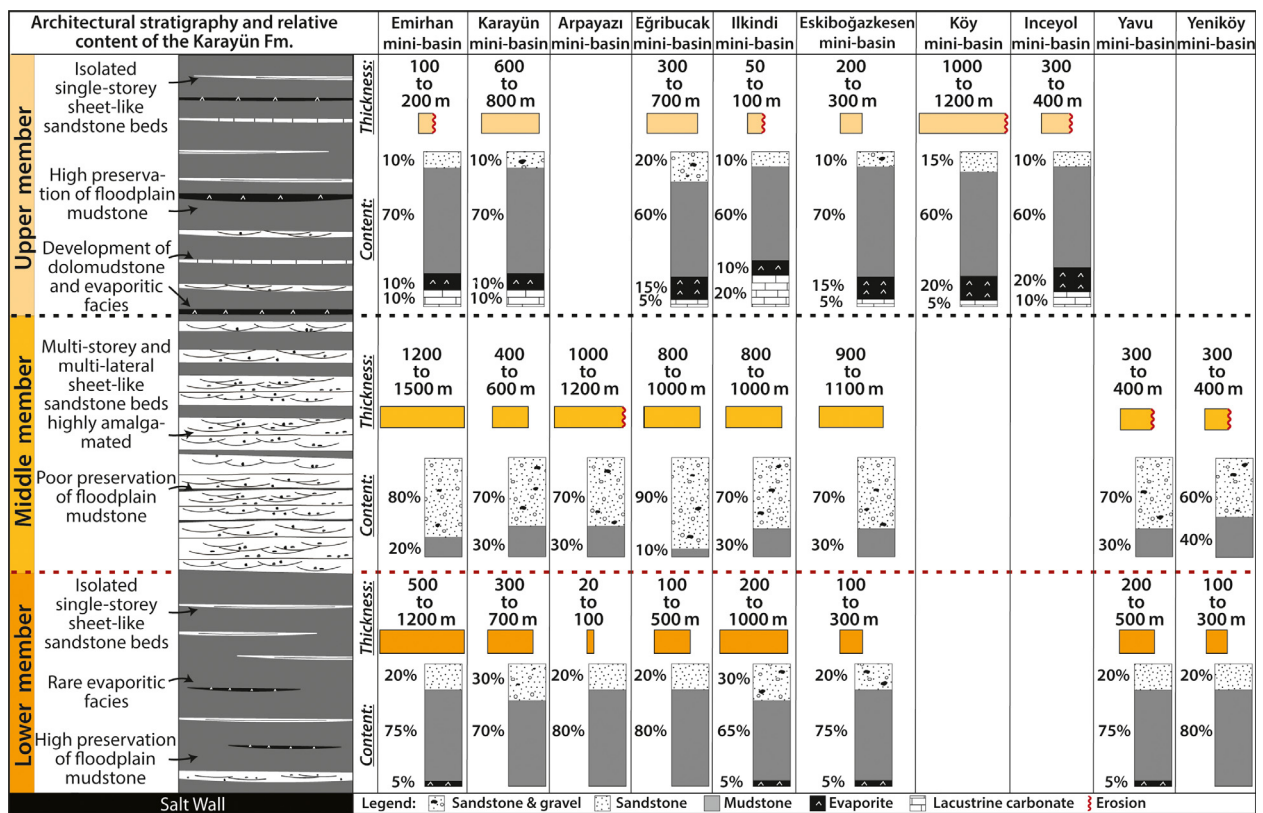
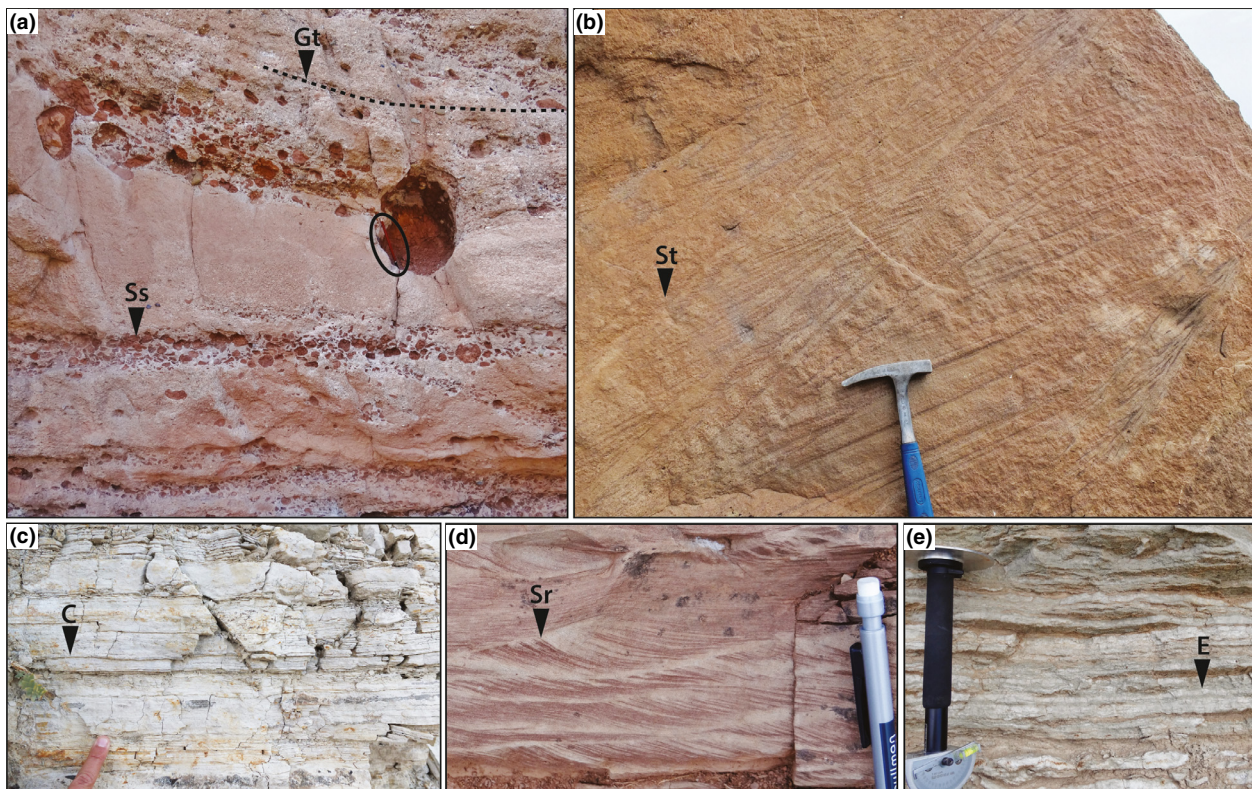


Fig. 5. Summary of the stacking patterns within the Lower, Middle and Upper members of the Karayün Formation in the studied mini-basins. Content and thickness variation of sand/conglomerate, evaporite, carbonate and floodplain mudstone within the Lower, Middle and Upper members in each mini-basin.





**Fig. 6.** Outcrop photographs of representative lithofacies in the Karayün Formation as observed in several mini-basins and summarized in Fig. 4. (a) Lithofacies Gt and Ss in the Middle member of the Arpayazi mini-basin, trough cross-stratified conglomerates (Gt) and scours filled by cobbles, granules and abundant intraformational mudclasts (Ss) representing a pebble channel lag. The intraformational clasts are sub-rounded to rounded and between 1 and 20 cm in size; pencil for scale: 0.10 m. (b) Lithofacies St in the Middle member of the Karayün mini-basin, medium-scale fluvial trough cross-bedding in fine- to coarse-grained sandstones; part of hammer for scale: 0.30 m. (c) Lithofacies C in the Upper member of the Ilkindi mini-basin, dark to blue grey carbonaceous micrites with ostracods and plants; part of hammer for scale: 0.3.0 m. (d) Lithofacies Sr in the Lower member of the Emirhan mini-basin, fine-grained to medium-grained sandstones with climbing-ripple lamination; part of pen for scale: 8 cm. (e) Lithofacies E in the Upper member of the Eğribucak mini-basin, evaporitic beds with laminar and fibrous forms (gypsum and anhydrite), part of Jacob staff for scale: 0.30 m. See Fig. 4 for explanation and interpretation of lithofacies.

trolled by raising and falling of the water table throughout deposition (e.g. Kraus & Aslan, 1999).

Thin evaporitic layers intercalated in red-green mudstones and associated fluvial sheet-sandstone bodies are interpreted as representing an ephemeral playa environment also called continental saline mudflats by Warren (2006). Gypsum and anhydrite were precipitated by sabkha-related processes in an extensive ephemeral saline lake (e.g. Warren, 2006). In contrast, the evaporitic clasts are interpreted as detritus derived from diapiric highs (e.g. Lawton & Buck, 2006). Emerging diapiric topography generated by salt withdrawal was eroded locally allowing reworking of diapir-derived detritus in the mini-basin (e.g. Lawton & Buck, 2006). However, the distribution of these evaporitic clasts is not restricted to the border of the existing diapir, so it is not possible to exclude a diapiric subaerial exposure upstream in the drainage area.

Facies assemblages making up the Lower member are interpreted as a system of terminal splay deposits on a floodplain with associated ephemeral sheet-floods interspersed with local development of continental saline mud-

flats environments (e.g. Warren, 2006; Nichols & Fisher, 2007).

### Transition between the lower and the middle members of the Karayün Formation

#### Description

The passage from the Lower to Middle member is expressed by an abrupt change from the reddish playa-lake floodplain muds with isolated sand bodies to sharp-based laterally extensive multi-storey fluvial sandstones and conglomerates of the Middle member (Figs 4, 5 and 7). This sharp surface marks the basal correlative surface of the Middle member expressed as a single vertically stacked sandy and conglomeratic bed that can be traced on aerial photographs.

#### Interpretation of faciological analysis and stratigraphic evolution

The upward transition between the playa-lake and the braided fluvial system is characterized by an abrupt basin-

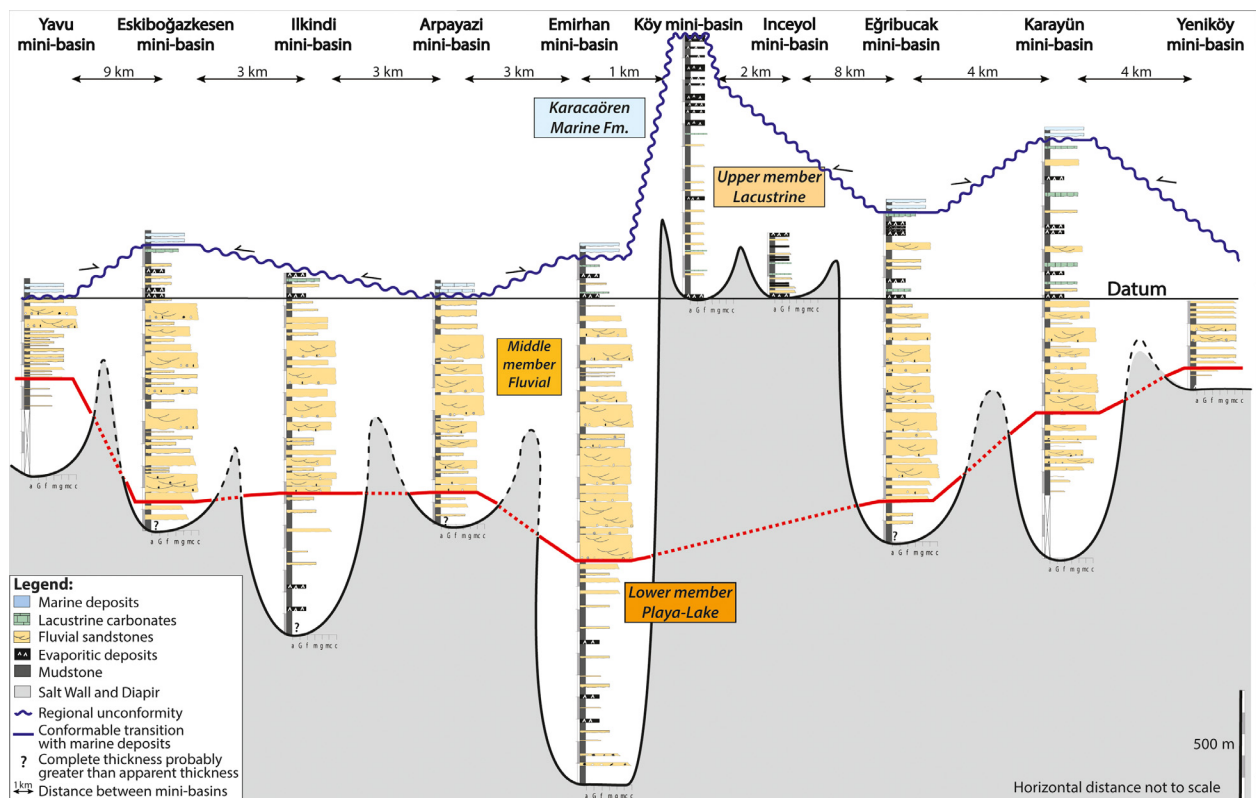


Fig. 7. Subregional SW-NE stratigraphic panel for the Karayün Formation based on field mapping and correlation of measured outcrop sections. The panel summarizes the main variations in thickness between the ten studied mini-basins. The first development of evaporitic facies is used as a correlative interval or datum for the basal limit of the Upper member. The measured sections are located in Fig. 3.

ward shift of facies, reflecting a significant decrease in the ratio accommodation/sediment supply (e.g. Martinsen *et al.*, 1999). This distinctive correlative surface corresponds to a sequence boundary (e.g. Martinsen *et al.*, 1999), which can be identified in all the studied mini-basins with a similar lithological and architectural expression (Figs 4, 5, 7 and 8b, c).

The abrupt renewal of coarser sedimentation everywhere in the studied area can be related to an uplift in the Teğer and Gürlevik areas to the south, considered as the sources of sediment supply (Temiz, 1996; Poisson *et al.*, 2010).

### Middle member of the Karayün Formation: fluvial braided deposits

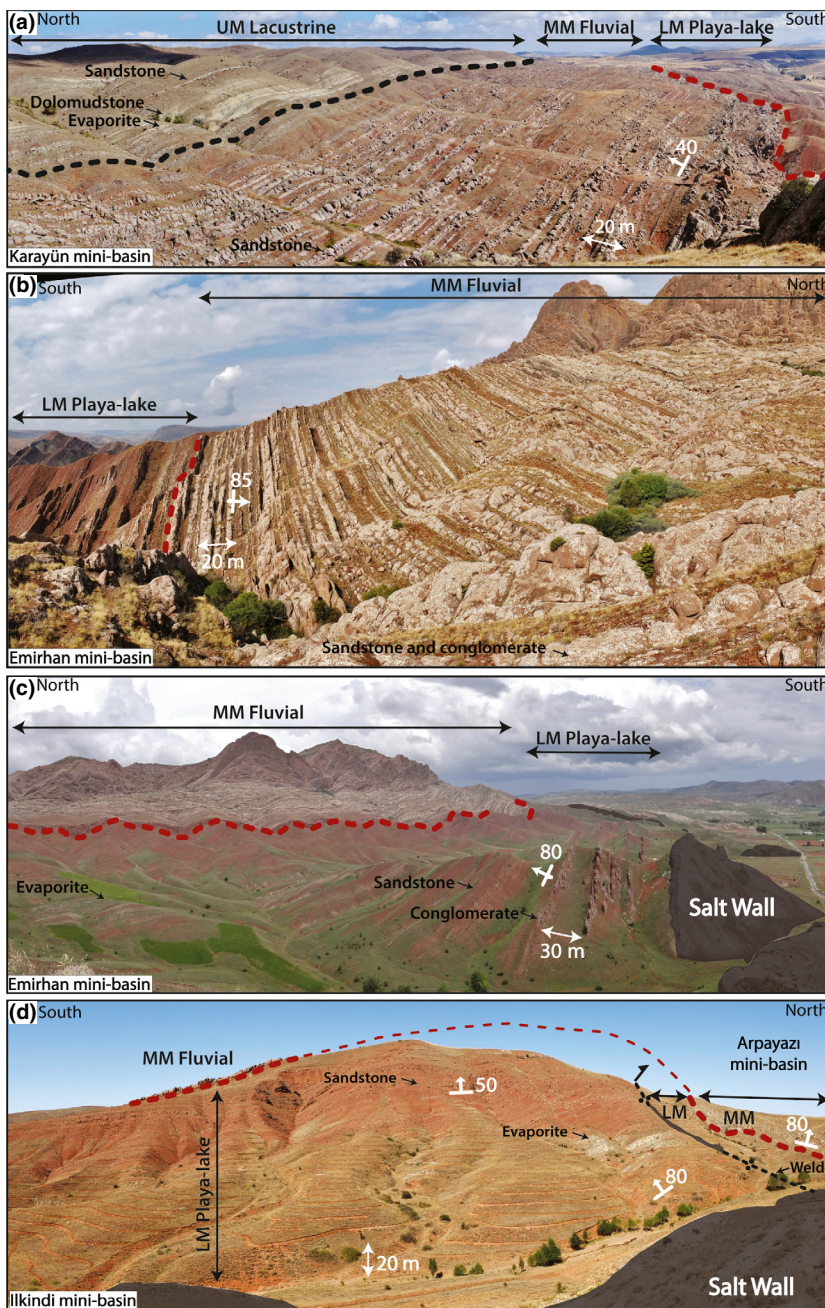
#### Description

The thickness of the Middle member varies from 200 to 1300 m in the studied mini-basins. Sandstone bodies form 5 to 40 m thick intervals within multi-lateral and vertically amalgamated channelized units showing high connectivity, comprising sandstones and conglomerates (60–90%) and mudstones (10–40%). Individual beds are 1–10 m thick and show fining-upward sequences with floodplain mudstone at the top. The sandstone bodies show mainly sheet-like geometries extending laterally over 2 km, with little evidence of large-scale incision.

These light brown and red sandstones are made up of medium-coarse grained deposits with well-sorted limestone, evaporite, serpentinite, peridotite and metamorphic clasts, locally containing red pebble-sized mud chips (Ss; Figs 4 and 6a). The sandstone beds are intercalated within laminated (Fl; Fig. 4) or massive (Fm; Fig. 4) red-oxidized mudstones, and can be traced laterally for more than 1000 m. The mudstone facies displays uncommon desiccation cracks and paleosol horizons composed of root traces with rhizoconcretions or white to grey calcrete nodules (Fp; Fig. 4).

The bases of the sandstone beds are characterized by a relatively well-scoured surface overlain by red intraformational mudstone pebbles and conglomerate clasts, generally 1–10 cm in diameter, aligned along the basal scour (Ss; Figs 4 and 6). This basal unit passes up into intervals with small- to medium-scale trough cross-stratification (St and Gt; Figs 4 and 6b), with a variety of dewatering structures, overlain by horizontally stratified sets (Sh) and, at the top, climbing-ripple and ripple-cross laminations (Sr). The latter lithofacies is picked out by red to light-grey mud-siltstone couplets, a few mm-thick. Sandy bedforms (SB) and gravel bars (GB) are the most commonly developed architectural elements, along with laminated sand sheets (LS; e.g. Miall, 1985).

The Middle member shows an overall thinning-upward trend expressing an upward decrease in channel amalgamation with thicker alluvial plain muddy intervals.



**Fig. 8.** Views of the Karayün Formation and the Lower, Middle and Upper members. The dashed red lines represent the contact between the Lower and Middle member. The dashed black lines represent the contact between the Middle and Upper member. Locations of panoramic views are shown in Fig. 3. (a) Transition between the Middle member (high proportion of fluvial conglomerate and floodplain mudstone) and the Upper member (saline lacustrine deposits comprising sandstones, evaporites, dolomudstones and floodplain mudstones) in the Karayün mini-basin. (b) Sharp transition between the Lower member (fluvial terminal splay and playa-lake deposits) and Middle member (highly amalgamated extensive sandstone bodies) in the Emirhan mini-basin. (c) Lower member in the Emirhan mini-basin comprising evaporitic, floodplain mudstone, sandstone and conglomeratic facies. (d) Lower member in the Ilkindi mini-basin comprising evaporitic, floodplain mudstone and sandstone facies.

*Interpretation*

Each individual depositional cycle suggests sedimentation during flash flood events (e.g. Stear, 1985; North & Davidson, 2012). The vertical succession, from internal scoured surfaces passing up into trough cross-bedding or horizontal bedding overlain by ripple lamination, records fluctuations in discharge during flood events (e.g. Stear, 1985). The abundant intraformational mudclasts and scours suggest erosion of mud plain facies. The common occurrence of small scale dewatering structures in the sandstone bodies indicates a high sedimentation rate (Plint *et al.*, 2001). The assemblage of sandy bedforms (SB), laminated sand sheets (LS), and gravel bars (GB) is characteristic of a low-sinuosity channel morphology (e.g.

Friend, 1977; Miall, 1996). The amalgamation of sandstone bodies, the lack of preservation of floodplain beds and the presence of red mudclasts at channel bases indicate that the channels reworked the adjacent floodplain deposits (Nichols & Fisher, 2007). The repeated migration of sandstone bodies may ‘cannibalize’ most of the floodplain, resulting in a relatively high proportion of sandstone (Shanley & McCabe, 1994). This highly variable discharge may be responsible for the poor development of incised channels and may ultimately facilitate a rapid avulsive pattern of channel relocation. The rapid aggradation of channels, the strong amalgamation and the increase in grain size can be related to more frequent episodic flood events accompanied by an intensification of overall discharge.

However, a limited tendency to channel migration is implied by the aggrading floodplain architecture, the dominance of vertically stacked channels and the aggradation of sandy bedforms in the floodplain deposits recorded by the uppermost part of this member (Friend, 1977; Miall, 1996). The sheet-like geometry of individual or composite sandstone bodies is interpreted as a braided system, where the interbedded red mudstones (FF) and thin sandstones (LS) represent the floodplain.

This facies association is characteristic of a proximal braided river plain, particularly in semi-arid regions where ephemeral runoff forms a network of shallow and poorly defined interlacing channels (e.g. Miall, 1996; Cain & Mountney, 2009; North & Davidson, 2012).

### Transition between the middle and the upper members of the Karayün Formation

#### *Description*

In the uppermost part of the Middle member, fluvial deposits show a change from multi-storey and multilateral sheet-like sandstones to single-storey units with an increase in the proportion of floodplain facies. The first occurrence of a saline lacustrine horizon marks the base of the Upper member (Fig. 4, 5 and 7).

#### *Interpretation of faciological analysis and stratigraphic evolution*

The accumulation of thick floodplain mudstones with a progressive transition from an amalgamated multi-storey to a single-storey fluvial architecture indicates an increase in the A/S ratio (accommodation space/sediment supply; e.g. Martinsen *et al.*, 1999). This change is recorded in all the mini-basins, supporting a regional control of the A/S ratio. The development of evaporitic facies is used as a correlative interval [or datum] corresponding to the basal limit of the Upper member (Figs 4, 7 and 8a).

### Upper member of the Karayün Formation: saline lacustrine deposits

#### *Description*

The upper member reaches a maximum thickness of 800 m. It is dominantly made up of floodplain deposits mostly composed of green to red mudstones (60–70%), with a few isolated fine- to coarse-grained sandstone bodies (10–20%) interbedded with lacustrine carbonate (5–20%) and evaporitic beds (10–20%; Figs 4, 5, 7 and 8a).

The evaporitic facies consists of 0.2–10 m thick beds of laminated gypsum or anhydrite, with a white and uniform macrocrystalline matrix. The most commonly observed structures are 'chicken wire fabric' mosaics forming closely packed anhydrite nodules 2–5 cm in diameter (E; Figs 4 and 6e).

The fine- to coarse-grained sandstones are grey to light brown in colour, forming sheets and lenticular bodies 0.2–2 m thick extending laterally for hundreds of metres and intimately interbedded with mudstone. Sandstone bodies occur as isolated single-storey units displaying a sheet-like geometry. The fine-grained sandstone intercalations show parallel laminations (Sl) and trough cross-beds (St), as well as small-scale symmetrical and a Sr laminations (Sr and Sw).

The limestone beds consist of beige to dark grey dolomitic mudstone, and attain thicknesses of a few dm (C; Figs 4 and 6c). These beds contain ostracods and plant debris such as charophytes. In the Upper member exposed in the Karayün, Eğribucak, Eskiboğazkesen, Ilkindi and Köy mini-basins, Poisson *et al.* (2012) recognised charophytes (*Lychnothamnus*) attributed to the late Oligocene.

#### *Interpretation*

This assemblage of facies with dark-grey dolomitic mudstone, red-green mudstone, evaporitic beds and sandstone bodies represents a progressive transition to a saline lacustrine environment (e.g. Dunagan & Turner, 2004; Warren, 2006). In contrast to the Lower member, which is dominated by an evaporitic playa-lake environment, the widespread occurrence of dolomitic mudstone facies reflects the establishment of a lake during deposition of the Upper member (e.g. Warren, 2006).

Sheet-like sandstones bodies are interpreted as sheet flood events across the flood-plain. The small-scale Srs in the sandstone-siltstone alternations are thought to be formed by wind-driven waves in water depths of a few dm (e.g. Allen, 1968). The evaporitic beds represent periods characterized by high evaporation rates relative to groundwater and surface water input into the ephemeral lake (e.g. Dunagan & Turner, 2004; Warren, 2006). This vertical alternation of sandy and carbonate beds can be attributed to high-frequency variations of lake-level through time.

### Palaeocurrent analysis and sediment provenance

Palaeocurrent directions were measured at 16 sites located throughout the Karayün Formation in all the studied mini-basins, yielding a range of dip directions from N200° to N80° with a vector mean by site ranging from N250° to N30° (Fig. 3). A dominantly NW-oriented direction of fluvial transport is recorded, parallel to the axis trend in most of the mini-basins (Fig. 3). No major changes in the palaeocurrent directions are recorded at the different sites from the base to the top of the Karayün section. However, some mini-basins such as Ilkindi, Eskiboğazkesen and Arpayazi display a large range of dip directions within the Middle member which are parallel to the mini-basin axis or towards the diapirs and salt walls.

The dominant north-westerly orientation of palaeocurrents observed at the regional scale indicates a southerly provenance for the clastic material supplying the Karayün Formation. This is consistent with the existence of outcrops further south comprising ophiolites of the neo-Tethyan suture, metamorphic and magmatic rocks of upper Cretaceous–Eocene volcanic rocks as well as limestones of the Teçer and Gürlevik Dağı (Fig. 3; Kurtman, 1973), all of which could have provided the clast types found in the Karayün Formation.

The scattering of palaeocurrents between the measurement sites is related to the structural behaviour of each individual mini-basin (i.e. local ponding) and the time of deposition. Moreover, the orientation of paleocurrent indicators were possibly modified by post-depositional rotation of some of the studied mini-basins. Also, the shape of the DFS drainage pattern is probably primary controlled by the salt topography (e.g. Banham & Mountney, 2013a; Venus *et al.*, 2015).

### **Towards a tectonic-forced depositional model: DFS**

The mapping and analysis of the three members suggest a fluvio-lacustrine system dominated by sheet flow processes during deposition of the Karayün Formation. The deposits of this system exhibit laterally extensive, non-erosional sand sheets interbedded with floodplain mudstones, sabkha-like evaporites and lacustrine limestones. Deposition of these sand-sheets involves brief pulses of rapid sedimentation on unconfined plains, with drastic fluctuations in water discharge and sediment transport. The occurrence of sheet-flood sandstones as well as playa-lake and lacustrine environments suggests that the basin-scale fluvio-lacustrine setting of the Karayün Formation was likely associated with a DFS spreading toward the North with a main entry point from the South (e.g. Nichols & Fisher, 2007; Hartley *et al.*, 2010; Weissmann *et al.*, 2010). DFSs are common in internally drained hydrologically closed basins, being governed by tectonic and climatic conditions (Weissmann *et al.*, 2013). DFSs typically exhibit proximal amalgamated channel belts with limited preservation of fine-grained floodplain material, less coarse material in intermediate zones within interspersed sheet-flood sandstone units, and distal terminal splay within playa-lake or lacustrine environments (Weissmann *et al.*, 2013). According to the Weissmann *et al.* (2010, 2013) classification, (i) the red oxidized playa-lake and sheet flood deposits dominantly present in the Lower member are interpreted as distal or basinal zones in a DFS model; (ii) the amalgamated fluvial Middle member is interpreted as the proximal zone of a DFS with proximal braid plain facies; and (iii) the saline lacustrine Upper member is interpreted as the distal part of a DFS in an ephemeral lake environment.

## **SALT TECTONICS FORCING ON SYN-TO POST-DEPOSITIONAL MINI-BASINS INFILL**

The tripartite infill of the Sivas Basin (i.e. Lower, Middle and Upper members) shows rapid changes in thickness from one mini-basin to another, with spectacularly well exposed stratal patterns expressed by wedges, erosional truncation and hook geometries along salt walls, welds or diapirs (e.g. Giles & Rowan, 2012; Rowan *et al.*, 2012). These thickness changes and the observed stratal geometries are interpreted as reflecting salt tectonic activity accompanying salt withdrawal/fall evacuation and salt diapir rise/inflation during sedimentation (e.g. Matthews *et al.*, 2007; Giles & Rowan, 2012). The observed stratal relationships indicate that deposition of the Karayün Formation was influenced at the local scale by the evacuation and inflation of salt diapirs bordering the studied mini-basins. The detailed effect of synsedimentary salt movement during the evolution and development of the mini-basins is discussed based on (i) the distribution of the facies assemblages, (ii) the stratigraphic thickness changes, and (iii) the stratal patterns discussed below.

### **Distribution of facies assemblages**

The relative proportion of facies recorded vertically and laterally within each mini-basin (playa-lake, braided fluvial system and lacustrine facies) not only vary from proximal to distal zones along the N–S axis of progradation of the DFS system, but also appear locally controlled by salt-induced topography.

#### *Lower member*

The infilling of a mini-basin seems to begin systematically with the early development of relatively small thin depocentres of dominantly flood plain shaly sedimentation, with fine-grained sheet-like sandstone bodies in most of the mini-basins. Only few conglomeratic beds are observed in the Karayün, Emirhan, Ilkindi and Eskiboğazkesen mini-basins (Fig. 9a). This early stage is followed by a period of coalescence of these small initial depocentres into larger individual depocentres occupying the centre of the mini-basins (Fig. 9b). The sedimentation remains dominated by floodplain mudstone and fine-grained sheet-like sandstone bodies in most of the mini-basins, with the occurrence of a basal depositional evaporitic interval in the Emirhan and East Ilkindi mini-basins, while the Karayün and West Ilkindi mini-basins appear persistently conglomeratic (Fig. 9b).

Within each mini-basin, coarse-grained sandstone and conglomerates tend to concentrate along the axis of the depocentres and pinch out laterally toward the diapir into more shaly facies.

However, the relative proportion of facies varies significantly between adjacent mini-basins. The Karayün, Egri-

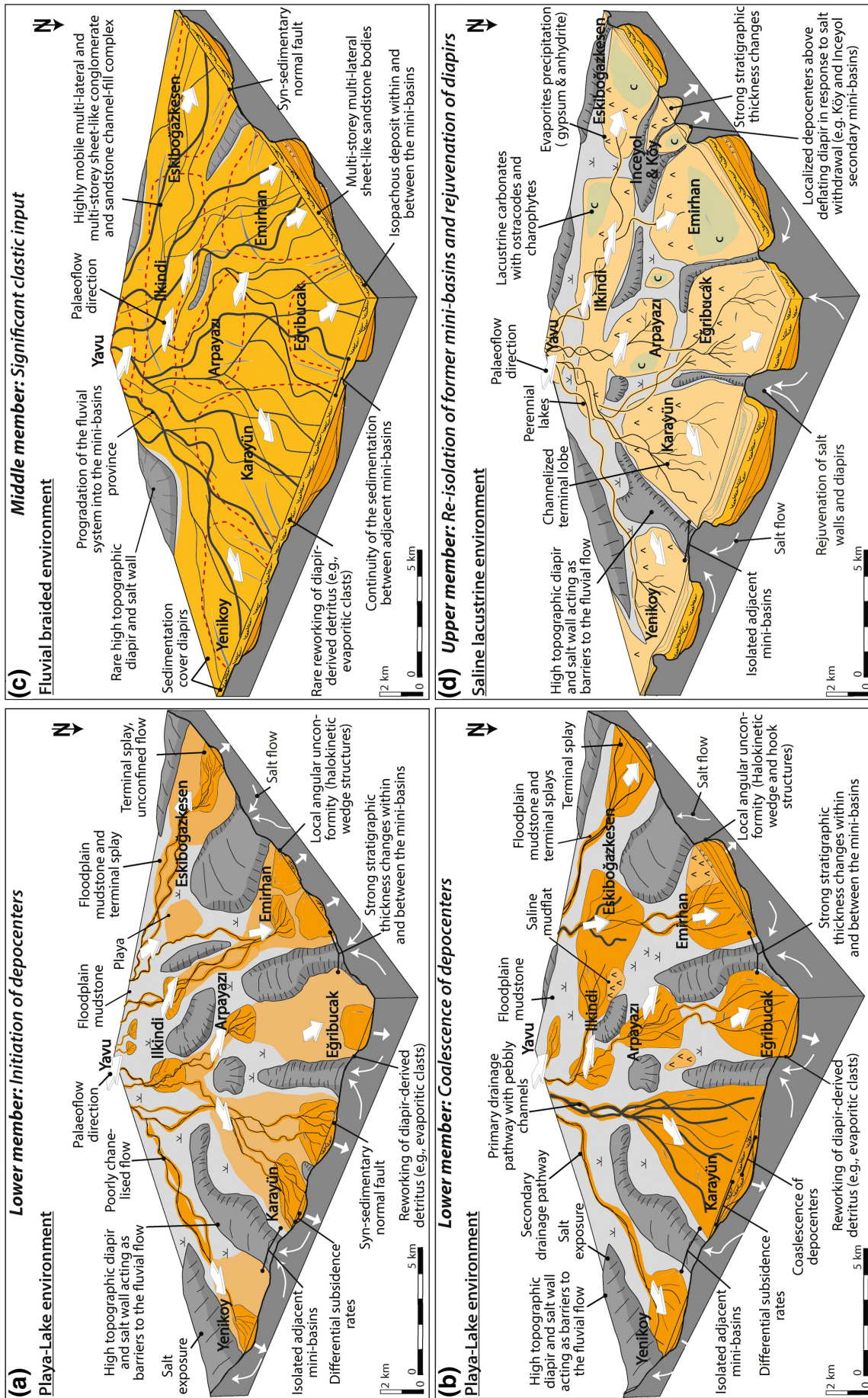
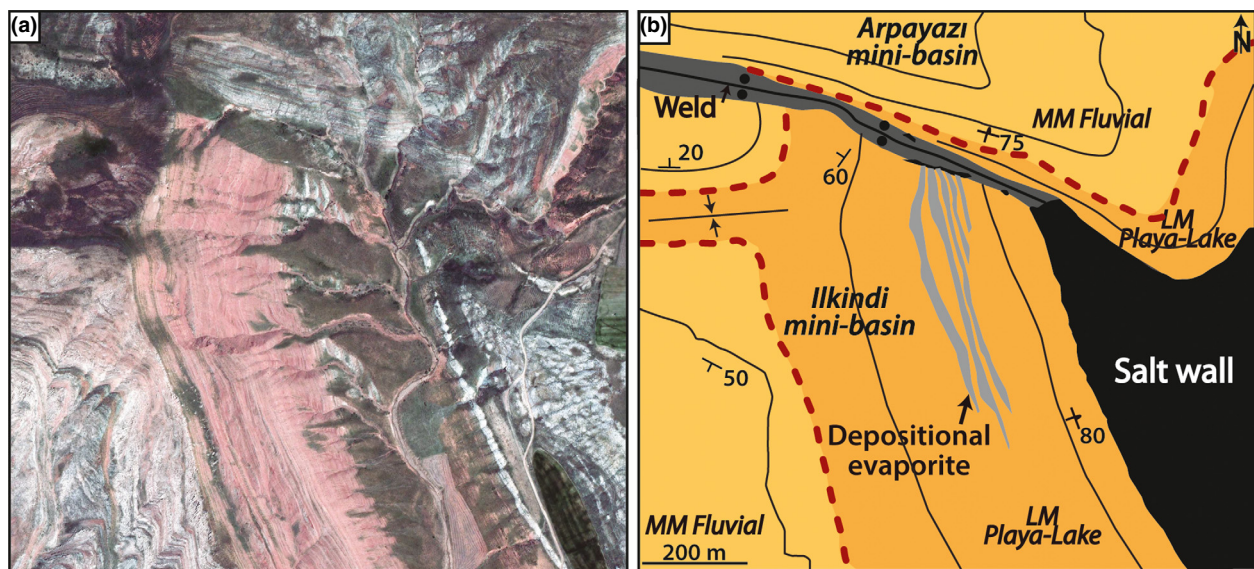


Fig. 9. Paleogeographic-tectono-stratigraphic diagrams of the Karayün formation in the studied mini-basins. (a) Depositional model illustrating sedimentary pattern of the basal part of the Lower member, showing the inception of a depocenter filled by small terminal splays. The sediment distribution across this area is influenced by elevated topographic highs produced by halokinesis. (b) Depositional model illustrating sedimentary pattern of the uppermost part of the Lower member, showing the coalescence of depocenters filled by terminal splays. The sediment distribution across this area is still influenced by elevated topographic highs produced by halokinesis. (c) Depositional model illustrating sedimentary pattern of the Middle member, showing invasion of voluminous clastic input into this area. This isopachous member seems to cover the diapir, and shows very little evidence of syn-sedimentary halokinesis. (d) Depositional model illustrating sedimentary pattern of the Upper member, showing establishment of a saline lacustrine environment associated with rejuvenation of diapirs leading to disconnection of the mini-basins.



**Fig. 10.** (a) Larger-scale version of the geoeye satellite colour composite image of Fig. 3, showing the northern border of the Ilkindi mini-basin. (b) Line drawing of the Ilkindi border showing depositional evaporitic beds in close contact with existing salt weld. These evaporitic beds tend to pinch out laterally a short distance away from the northern weld and can be interpreted as chemically remobilized evaporites formed locally in the playa-lake environment of the Lower member. The dashed red lines represent the contact between the Lower and Middle member.

buçak and the western part of the Ilkindi mini-basins appear more prone to sandy sedimentation throughout Lower member (Fig. 9a, b), in contrast to the Emirhan and the eastern part of the Ilkindi mini-basins which record a relatively sand-starved vertical evolution with evaporitic beds interbedded with floodplain mudstone facies (Fig. 8c, d). In the southern mini-basins (Eskiboğazkesen, Yavu and Yeniköy), the sedimentation is dominantly shaly with no evaporites.

The abundance of conglomerates found in the Karayün mini-basin, as well as in the neighbouring Egribuçak and western part of the Ilkindi mini-basins, is interpreted as the expression of the main transit axis of the main regional DFS fan (Fig. 9a, b). On the other hand, mini-basins such as Yeniköy, Yavu and Eskiboğazkesen are located along secondary drainage pathways characterized by fine-grained terminal splay deposits (Fig. 9a, b). In the Emirhan and eastern part of the Ilkindi mini-basins, the youngest mini-basins infills are thought to be isolated from the main drainage axis, with dominant floodplain mudstone and evaporitic deposits (Fig. 9a, b).

Differential subsidence induced by salt evacuation, and possibly influenced by the inherited structural pattern, creates local accommodation space in the centre of mini-basins that concentrate the coarse-grained sedimentary input. The behaviour of subsiding depocentres linked to a salt withdrawal in specific areas is not only expressed by facies assemblages but also by halokinetic growth strata (i.e. wedge structure). Furthermore, the particular distribution of the facies assemblages between the adjacent mini-basins demonstrates the confinement induced by topographically elevated salt walls or diapirs (Fig. 9a, b). The diapiric topography generated by salt withdrawal

segregates and diverts the fluvial pathways, and thus locally controls the sediment dispersal and facies pattern (e.g. Banham & Mountney, 2013a; Venus *et al.*, 2015). The topographically elevated exposed evaporite material likely represents a source for the remobilization of sulphate-rich fluids feeding the playa-lake environment, thus facilitating the precipitation of evaporitic beds in the mini-basins (e.g. Lawton & Buck, 2006; Paz & Rossetti, 2006). In addition, some depositional evaporitic beds in close contact with existing salt welds tend to peter out laterally at a short distance from these welds and can be interpreted as due to local chemical remobilization of evaporites (Figs 8d and 10).

#### *Middle member*

The Middle member is dominated by fluvial coarse-grained channelized related deposits. It is relatively homogeneous in term of facies distribution and little thickness variation is observed within each mini-basin (Fig. 8a, b). Some subtle facies changes are noted in relation to large-scale proximal to distal evolution of the fluvial system, but without any compartmentalization of the mini-basin sediment input.

The development of the fluvial system during this time interval is regarded as unconfined and poorly constrained by salt-induced topography, being associated with vertical stacking of the long-lived depositional system (Fig. 9c).

#### *Upper member*

The Upper member lies conformably on top of the Middle member in five mini-basins (Eskiboğazkesen, Ilkindi,

Emirhan, Karayün and Eđribucak), or forms individual small mini-basins (Figs 3, 5 and 7). The Upper member was probably eroded by recent uplift and erosion in the Yavu, Yeniköy and Arpayazi mini-basins. In the Miocene marine unconformable bed capping the Arpayazi mini-basin, the presence of angular blocks of lacustrine carbonate may indicate an early erosion and re-sedimentation of this lacustrine deposit into the Miocene section prior to recent erosion. In the small Inceyol and Köy mini-basins, the Upper member seems to overlap directly on top of collapsed salt diapirs (Fig. 3).

The Upper member can be relatively well recognized regionally and appears homogenous in terms of facies assemblages. However, some noticeable variations are observed (Fig. 5). In the Karayün and Eđribucak mini-basins, a higher amount of fine sand is preserved compared to other mini-basins, possibly indicating a preferential drainage axis straddling these two mini-basins (Figs 5, 7 and 9d). In these two mini-basins, thicker and wider evaporitic and lacustrine intervals are also recorded.

Farther south-west, the Upper member in the Eskiboğazkesen mini-basin is developed with some lacustrine mouth bar deposits indicating a possible secondary drainage axis (Figs 5 and 7). In the Inceyol and Köy mini-basins, and in the small remaining parts of sections near the top of the neighbouring Emirhan and Ilkindi mini-basins, only floodplain mudstone evaporites and limestones are present. This could indicate the development of starved mini-basins which are nevertheless difficult to understand precisely due to the limited size of the preserved exposures.

As observed in the Lower member, lateral facies variations across the mini-basins during accumulation of the Upper member indicate a renewed influence of salt topography on depositional trends (Fig. 9d).

### Regional isopach patterns and local thickness variations

The overall Karayün Formation thickens along a West-East axis in the central part of the Sivas Basin, with an estimated maximum thickness of more than 3500 m (Fig. 7). The Emirhan mini-basin shows the thickest preserved continuous section (2400 m). Due to recent erosion and uplift, it is difficult to draw up an overall isopach map for the whole region. However, some noticeable thickness variations can be pointed out within the Karayün Formation.

#### Lower member

The Lower member displays pronounced thickness variability within and between the neighbouring mini-basins (Figs 5 and 7). The thickest section is in the Emirhan mini-basin, reaching 1200 m at a maximum and thinning laterally to 400 m in the eastern part of the mini-basin (Figs 3 and 5). Compared to the Emirhan mini-basin, the

Lower member in the Arpayazi mini-basin reaches a maximum thickness of 100 m (Figs 3 and 5). The thickness of the Lower member section in other adjacent mini-basins varies between 100 and 1000 m (Figs 3 and 5).

The thickness variations within and between mini-basins reflect the initial distribution of depocentres created by salt withdrawal. However, some mini-basins are only partially tilted showing probably only the edge of the Lower member at outcrop and exhibiting only apparent thicknesses (Figs 2 and 7).

#### Middle member

The Middle member is present as a broad relatively isopachous (900–1200 m) interval in the central part of the Sivas Basin, with thinning toward the border of the studied area (Figs 5 and 7). The Middle member is 900 m thick (i.e., Eskiboğazkesen, Ilkindi, Arpayazi and Eđribucak), being slightly thicker (1200 m) in the Emirhan mini-basin and thinner (600 m) in the Karayün mini-basin (Figs 5 and 7).

Furthermore, an enlargement of the fluvial system is observed towards the salt walls and diapirs, clearly observed in the Eđribucak, Eskiboğazkesen and Yavu mini-basins (Figs 7 and 8c). During deposition of the Middle member, the fluvial system was unconfined by surrounding diapirs as these were likely buried with relatively minor topographic expression and/or differential subsidence (Fig. 9c). The marked enlargement of the fluvial system within a regionally isopachous deposit indicates an abrupt increase in the ratio between sediment accumulation rate and diapiric salt-rise (Figs 8c; e.g. Giles & Rowan, 2012). The isopachous character of the Middle member indicates that local accommodation created by salt diapir withdrawal was overwhelmed by sediment supply, both in previously developed mini-basins and over former diapirs.

#### Upper member

The Upper member fully preserved in the Eskiboğazkesen (300 m thick), Eđribucak (700 m thick) and Karayün (800 m-thick) mini-basins displays some marked thickness variations (Figs 5 and 7). Intra mini-basins thickness variations are limited and expressed by a relative small thinning toward the mini-basins borders (Figs 5 and 7). In the particular setting of the Inceyol and Köy mini-basins, which directly overlie a present-day salt diapir, a maximum thickness of 1200 m is observed (Fig. 7).

The development of a very thick depocentre, as observed in the Köy and Inceyol mini-basins, could be interpreted as encapsulated or secondary mini-basins. These depocentres formed above an original salt feeder crest showing a progressive downbuilding of the mini-basin, leading to isolation and tightening of the syncline (e.g. Smith *et al.*, 1993). The development of these secondary mini-basins is interpreted as due to salt with-



drawal directly related to the local accommodation generated by salt evacuation.

However, the constant thickness across the Eskiboğazkesen, Eğribucak and Karayün mini-basins prevents local accommodation directly related to mini-basin downbuilding (Fig. 9d).

The Upper member in the Emirhan mini-basin was also eroded during the Early Miocene, while later erosion was enhanced by rotation and tilting of the mini-basin.

### Stratal geometries associated with salt-related deformation

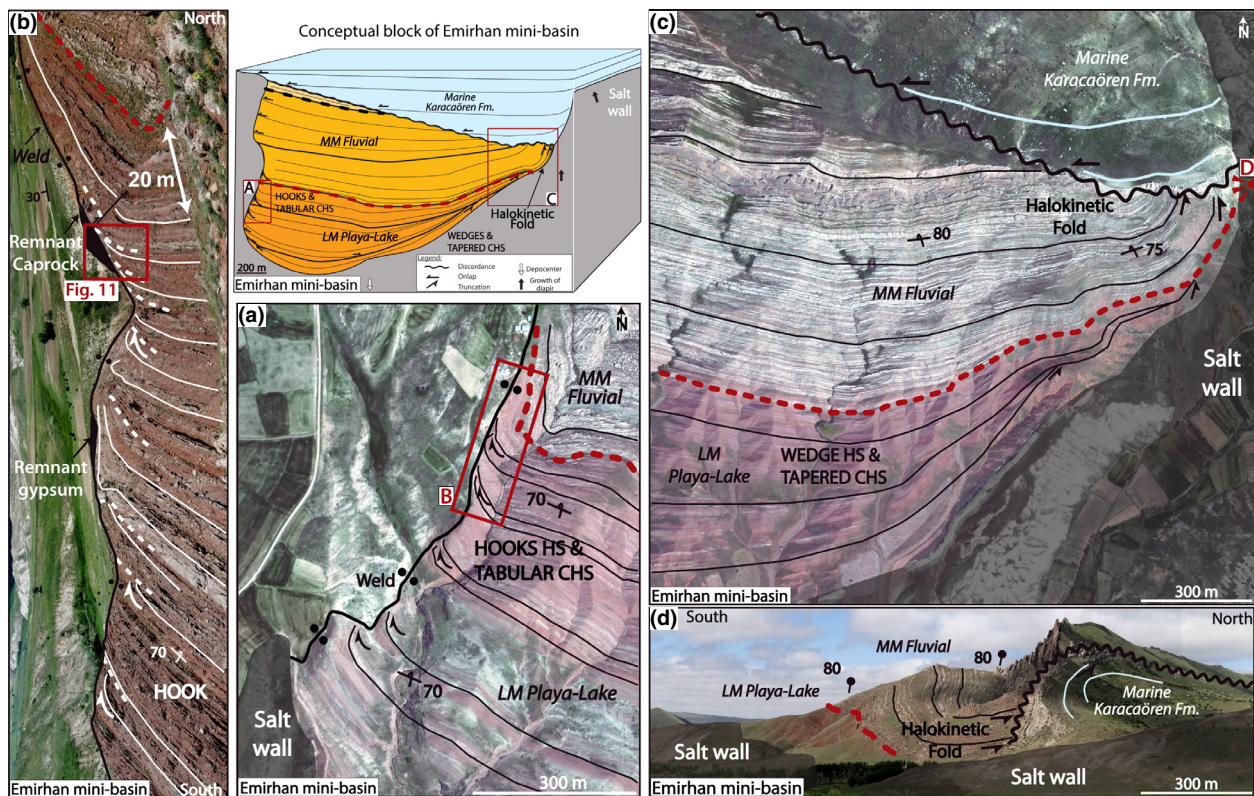
Apart from the thickness variations discussed above, different types of halokinetic stratal patterns flanking the borders of the mini-basins are described that record different periods of relative salt-structure growth. Three basic types of halokinetic stratal geometry and two associated composites halokinetic sequences were defined by Giles & Rowan (2012), and are also recognized in the Sivas Basin:

(1) Small-scale hook-shaped folds are characterized by narrow and steep drape-fold geometries very close to

the mini-basin border. Beds are folded and thin out over a short distance from the diapir, with a bed rotation of up to 90°. Bounding unconformities are correspondingly highly angular with up to 90° of angular discordance, but become conformable within a few metres of the diapir. The vertically stacking of hook geometry creates a large-scale package with tabular form, named tabular composite halokinetic sequences (tabular CHS).

(2) Small- to medium-scale wedge structures are characterized by broad and gentle drape-fold geometries. Folding and thinning of the growth strata occur over a wider zone of 100–1000 m, with only minor and gradual bed rotation. Bounding unconformities have low-angle truncation, but commonly extend for more than 500 m away from the diapir before becoming conformable. The vertical stacking of wedge geometry creates a large-scale package with a broadly folded tapered form, named tapered composite halokinetic sequences (tapered CHS).

(3) Large halokinetic folds are characterized by steeply dipping strata draped onto the diapir flanks over a large vertical interval. Strata in these large halokinetic folds may be of constant thickness or thin



**Fig. 11.** Conceptual block diagram showing the Emirhan mini-basin. The dashed red lines represent the contact between the Lower and Middle member. The dashed black lines represent the contact between the Middle and Upper member. (a) Larger-scale version of Fig. 3b illustrating a succession of hook-shaped stratal geometries and the tabular composite halokinetic sequence on the western border of the Emirhan mini-basin Lower member. (b) Photo view of a typical succession of small hook structures affecting the playalake Lower member. (c) Larger-scale version of Fig. 3b illustrating the wedge-shaped stratal geometry and the tapered composite halokinetic sequence of the eastern border of the Emirhan mini-basin Lower member. (d) Photo view of a large hook structure affecting the fluvial Middle member, truncated and overlapped by marine deposits.

towards the diapir, whether it is preserved or truncated.

Erosional truncation on top of the rotated strata are observed between the individual halokinetic hooks, wedges or large folds, generating a surface that is subsequently overlapped and overlapped, producing a localized angular unconformity.

Lower member

The best examples of small- to medium-scale wedges and composite tapered CHS are seen in the Lower member of the Emirhan and Karayün and Ilkindi mini-basins (Fig. 7). Near the base, the observed wedges form onlap geometries on top of the salt (Figs 9a, 11 and 12). With migration of the depocentre, several examples of intraformational wedges are seen higher up in the sections, being particularly well expressed in the centre of the Karayün mini-basin (Fig. 12). Several small wedge structures are also observed in the western Ilkindi, eastern Emirhan and northern Eskiboğazkesen mini-basins (Fig. 11c). The

Lower member near the western border of the Karayün mini-basin exhibits a succession of broad and gentle drape-fold geometries within 70–100-m-thick wedges of conglomerates with associated erosional truncation developed near the salt wall (Fig. 12a). Folding and thinning occur over a wider zone of 100–300 m, with only minor and gradual bed rotation. Bounding unconformities have low-angle truncation ( $>30^\circ$ ), but commonly extend for more than 200 m away from the border of the stratigraphic Unit before becoming conformable.

At a smaller scale compared to the wedges, individual depositional sequences, 10–100 m thick, show some stratal terminations with narrow and steep drape-fold geometries (i.e. hook shaped). Indeed, several small-scale hook structures affecting individual sequences and associated tabular CHS are observed in the western Emirhan (Fig. 11a, b), the eastern Ilkindi and the Eğribucak mini-basins. These hook structures display folding and thinning of the beds over a distance of 10–50 m from the diapir, with bed rotation of up to  $90^\circ$  and rare overfolding. Bounding unconformities are of erosional truncation type

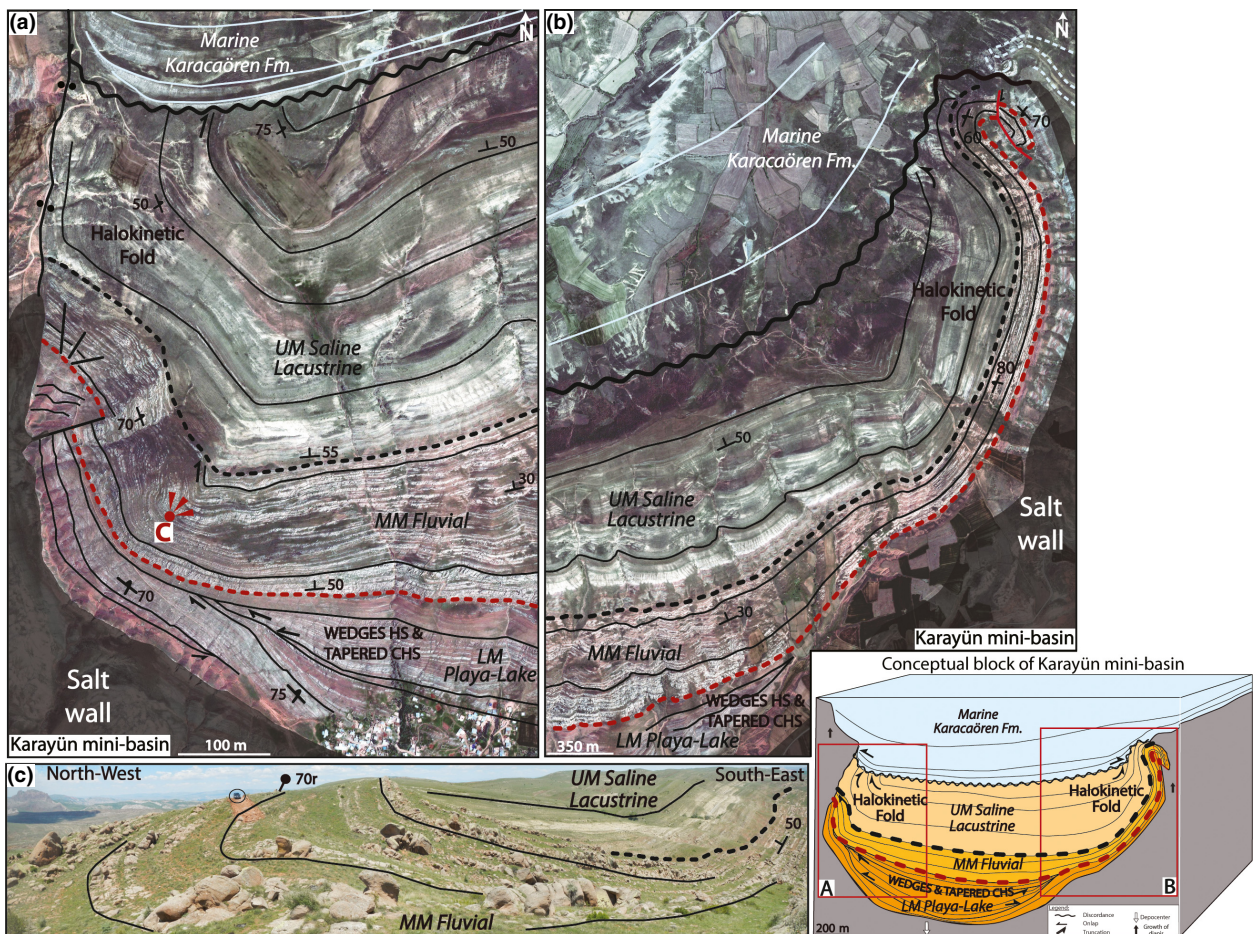
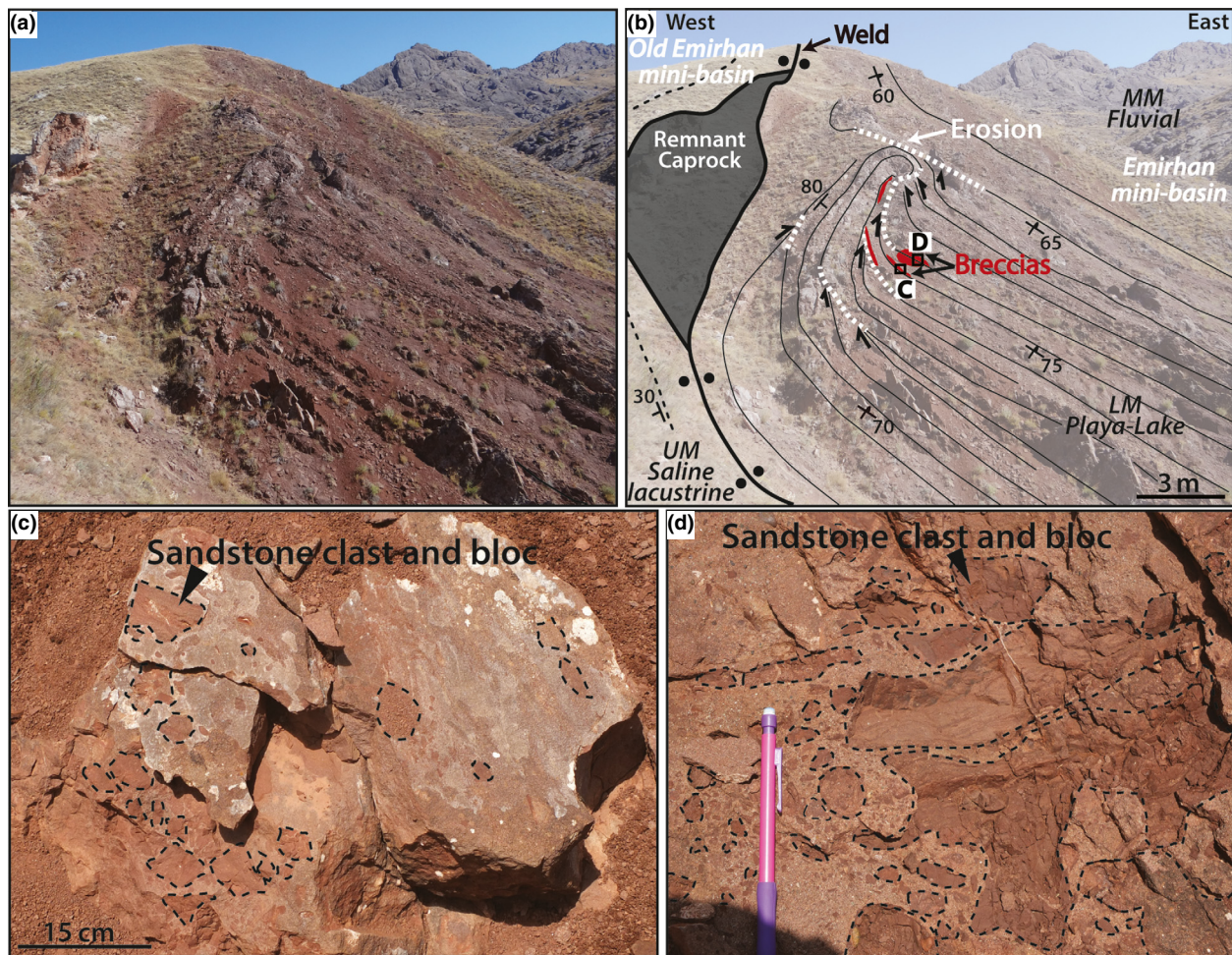


Fig. 12. Conceptual block diagram showing the Karayün mini-basin. The dashed red lines represent the contact between the Lower and Middle member. The dashed black lines represent the contact between the Middle and Upper member. (a) Larger-scale version of Fig. 3b illustrating stratal pattern of the western border of the Karayün mini-basin western boundary. (b) Larger-scale version of Fig. 3b illustrating stratal pattern of the eastern border of the Karayün mini-basin. (c) Photo view displaying folding of the fluvial Middle member on the western boundary.



**Fig. 13.** Outcrop photographs illustrating the geometry of the western border of the Emirhan mini-basin. (a) View of individual hook structures. (b) Line drawing of detailed hook structures. (c and d) Views of destabilisation breccias derived from the upturned underlying strata. This facies consists of matrix-supported fine to coarse-grained sediment containing angular pebbles or cobbles of the underlying strata.

corresponding to high-angle discordances ( $<90^\circ$ ), but become conformable within 50 m of the diapir (Fig. 11b). In each small sequence, there are no variations in thickness or sand content across the hook structures (Fig. 11a, b). However, within individual hook structures, localized destabilisation breccias are observed derived from the upturned underlying strata (Fig. 13). This facies consists of matrix-supported fine to coarse-grained sediment with angular pebbles or cobbles of the underlying strata (Fig. 13c, d). These destabilisation products are formed during the folding and make up small accumulations of limited extent from 0.5 to 2 m (Fig. 13b). A slight migration is recorded though time, within a general drift toward the mini-basin centre, but these facies typically do not extend more than a few metres away from the diapir (Fig. 13b).

These stratal relationships including small hooks and small to medium wedge structures express a continuous process of salt evacuation during downbuilding of the mini-basins, with a low sediment-accumulation rate relative to diapiric salt rise on the time scale of deposition of

the Lower member (e.g. Giles & Rowan, 2012). The development of the mini-basins during the deposition of the Lower member is clearly influenced by salt-induced local accommodation generated by salt withdrawal (Fig. 9a, b).

Furthermore, despite an approximately constant rate of sedimentation accumulation within the Lower member, different halokinetic sequence and composite halokinetic sequence geometries are observed in time-equivalent strata on opposite sides in the Emirhan mini-basin (Fig. 11). This asymmetric mini-basin fill suggest different rate of salt rise between both flanking diapirs.

#### *Middle member*

Only large halokinetic folds are visible in the Middle member, affecting in particular the 300 m basal interval in the Emirhan, Yavu and Arpayazi mini-basins (Figs 11c, d and 14). Along the eastern border of the Emirhan mini-basin, a large fold is seen forming a large

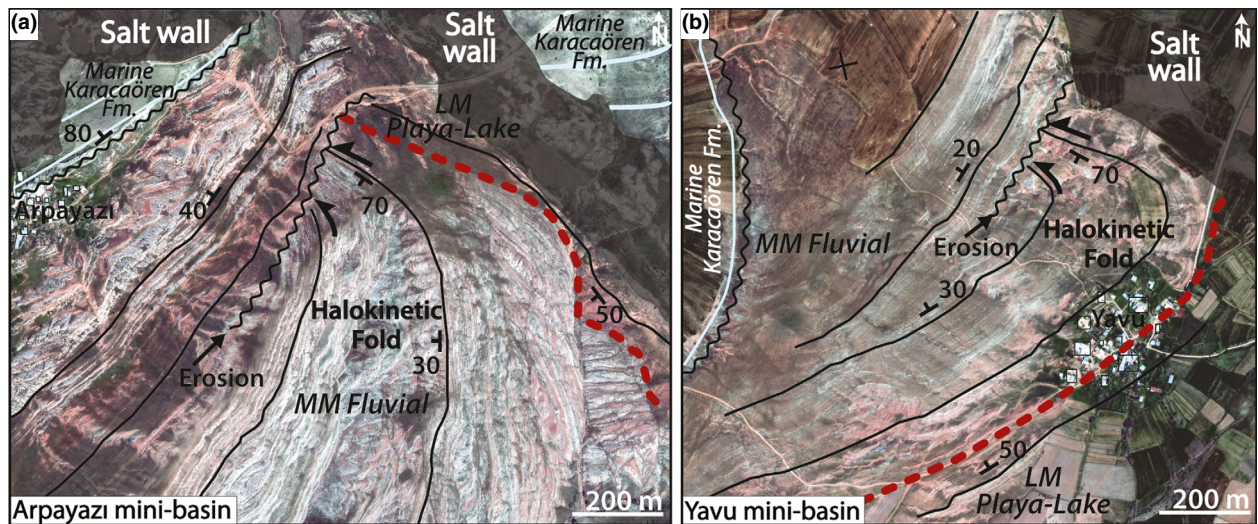


Fig. 14. (a) Larger-scale version of Fig. 3 illustrating the hook-shaped stratal geometry of the Arpayazi mini-basin in the fluvial Middle member. (b) Larger-scale version of Fig. 3 illustrating hook-shaped stratal geometry of the Yavu mini-basin in the fluvial Middle member. The dashed red lines represent the contact between the Lower and Middle member.

upturned group of strata which is truncated and overlapped by the marine Karacaören Formation (Fig. 11c, d). In the Yavu and Arpayazi mini-basins, the halokinetic folds are overlain by the upper part of the Middle member (Fig. 14).

On the western border of the Karayün mini-basin, the Middle member is upturned after deposition and partially eroded in the vicinity of the salt wall and overlapped by the Upper member (Fig. 12a). On the eastern flank, near-vertical strata of the Middle member is upturned over a distance of about 2 km adjacent to the eastern diapir. Rotation of strata off the top of the growing diapir and ongoing minibasin subsidence resulted in the preservation of strata draped along the steep flank of the diapir for up to 2 km (Fig. 12b, c). The sediment in this structure shows no variation in sand content (Fig. 12b, c).

The large halokinetic folds affecting the homogeneous and isopachous Middle member suggest widespread capping of diapirs and salt walls, followed by post-depositional folding induced by a regional increase in the vertical diapiric flow (Fig. 9c).

The coeval development of large-scale halokinetic deformation in numerous mini-basins suggests a regional control affecting the net diapir-rise rate (Figs 11c and 14). This halokinetic deformation is attributed to a rejuvenation of diapir ascent primarily influenced by shortening (C. Kergaravat, J.-P. Callot, C. Ribes & J.C. Ringenbach, submitted). The shortening seems to accentuate the salt flow, amplifying the halokinesis and thus influencing the development of mini-basins by diapir rejuvenation (Vendeville *et al.*, 1993; Vendeville & Nilsen, 1995; Davison *et al.*, 2000; Rowan *et al.*, 2003; Trocmé *et al.*, 2011; Callot *et al.*, 2012; Ferrer *et al.*, 2012; Saura *et al.*, 2014). In addition, considering the abrupt influx of fluvial sediment, the diapir-rise rate change could also represent a slightly delayed response to the blanket sediment loading.

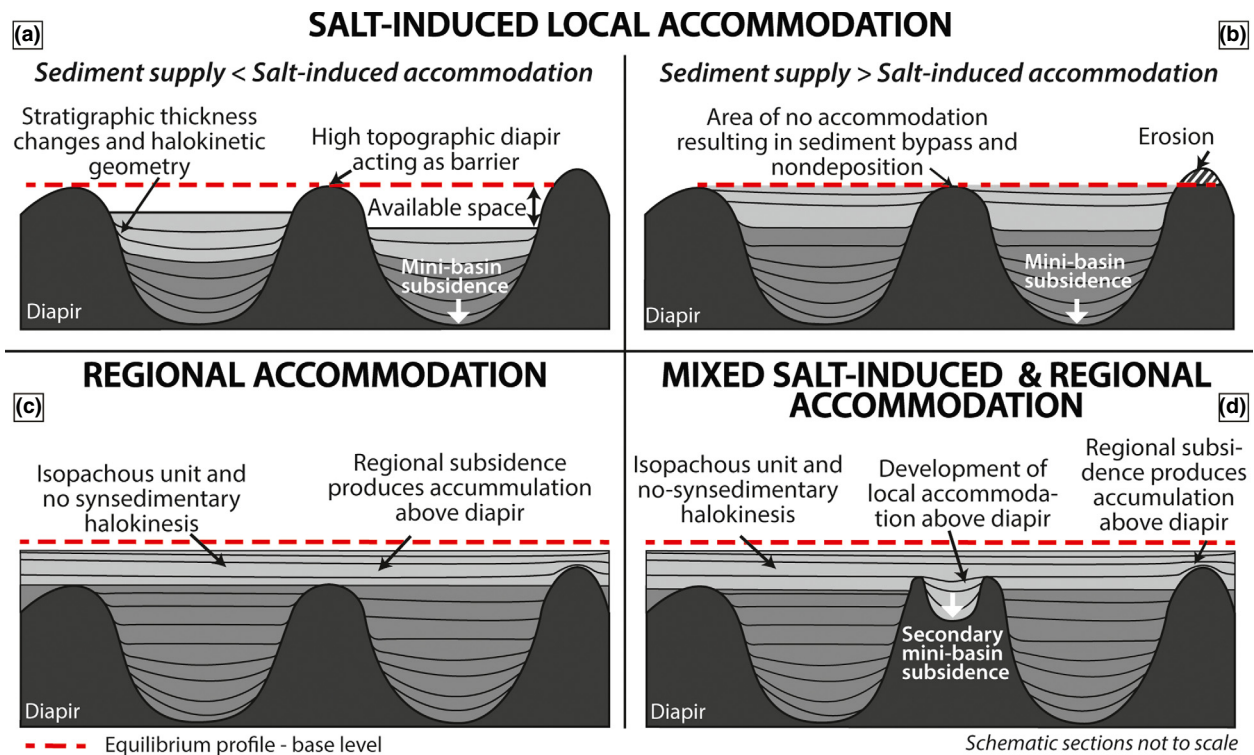
#### Upper member

The Upper member displays thick wedges in the Köy and Inceyol secondary mini-basins (Fig. 3). In contrast, the Eskiboğazkesen, Eğribucak and Karayün mini-basins display only very gentle wedging, implying limited salt withdrawal during the deposition of the Upper member (Figs 3 and 5). In the Karayün mini-basin, the overall member forms a wedge upturned on both sides (Fig. 12). The western border of the Karayün mini-basin records a stratal rotation near the western salt wall, with an erosional truncation corresponding to high-angle discordance up to 90° (Fig. 12a). However, the eastern border of the Karayün mini-basin exhibits an erosional truncation with 50–60° angular discordance (Fig. 12a). This erosional truncation is overlapped by the lowermost marine bed of the Miocene Karacaören Formation (Fig. 12a).

The slight wedging and presence of homogeneous facies assemblages across the Karayün, Eskiboğazkesen and Eğribucak mini-basins imply a capping of salt walls followed by post-depositional deformation of strata in the Upper member of the Karayün mini-basin (Fig. 9d).

## DISCUSSION

During the entire Oligocene, fluvio-lacustrine sediments accumulated in the Sivas Basin with no evidence of any marine connection (Poisson *et al.*, 2011). The first and unique marine incursion following the Late Eocene retreat and deposition of the thick mother salt is represented by the marine Karacaören Formation dated as Aquitanian (Kurtman, 1973; Poisson *et al.*, 1996). At that time, the Sivas Basin formed an elongated endorheic Basin isolated from the Paratethyan domain to the East (Poisson *et al.*, 1996, 2011). Therefore, during the Oligo-



**Fig. 15.** Four conceptual sketch sections showing the role played by halokinesis in the generation of regional accommodation. (a) Salt-induced local accommodation with low sediment supply resulting from mini-basin subsidence induces changes in stratigraphic thickness which accompany the development of growth strata and small halokinetic structures flanking the mini-basin borders, such as during deposition of the playa-lake Lower member. The topographically elevated salt walls and diapirs are especially well expressed when the sediment supply cannot fill the available accommodation and thus exerts a tight control on the sediment dispersal pattern. (b) Salt-induced local accommodation with high sediment supply resulting from mini-basin subsidence induces stratigraphic thickness changes accompanying the development of growth strata and small halokinetic structures flanking the mini-basin borders such as during deposition of the playa-lake Lower member. When the sediment supply is able to fill the available accommodation produced by mini-basin subsidence, the salt walls and diapirs are not topographically elevated and thus do not exert a local control on the sediment dispersal, such as in the case of the Black Ledge unit of the Chinle formation in the northeast Paradox Basin. (c) Tectonically driven regional accommodation attributed to the foreland flexure leads to the deposition of relatively isopachous series draping and finally obscuring the salt topography, such as observed in the fluvial braided Middle member. (d) Regional and coeval salt-induced accommodation produce very small changes in thickness and facies across the mini-basins, but secondary mini-basins can develop related to local subsidence on top of diapirs driven by salt movement at depth, such as during deposition of the Upper member of the Karayün Formation.

cene, sedimentation and stratigraphic preservation or sediment bypass and erosion are dependent on local base level rather than sea level fluctuations (e.g. Wheeler, 1964; Shanley & McCabe, 1994). The recorded sedimentary cycles within the Karayün Formation are controlled by a range of allocyclic factors, including tectonically driven regional subsidence, sediment supply, climate impact and tectonic setting.

The concept of base level profile is used to characterize the forcing parameters which control depositional sequences. The base level profile, corresponding to the upper limit of the accommodation space for sediment accumulation (Blum & Törnqvist, 2000), is determined by the balance between tectonic uplift of the hinterland (wedge top) and the correlative subsidence of the foreland (e.g. Fisher & Gary, 2013). Here, two distinctive scales and mode of subsidence are recorded in the Karayün Formation, and have significant implications

on the overall thickness and stratal geometry in the studied mini-basins:

- Tectonically driven regional accommodation attributed to the foreland flexure (Poisson *et al.*, 1996) induced the deposition of relatively isopachous series draping and finally obscuring the salt topography, such as developed in the fluvial braided setting of the Middle member (Fig. 15c).
- Salt-induced local accommodation resulting from salt withdrawal and diapir growth produced stratigraphic thickness changes and small-scale halokinetic structures along the mini-basin borders, such as developed in the playa-lake setting of the Lower member (Fig. 15a, b).

However, regional and salt-induced accommodation can coexist, such as during the deposition of the Upper member of the Karayün Formation (Figs 9d and 15d). In this

case, very small thickness and facies changes are recorded across the mini-basins, but secondary mini-basins can develop related to local subsidence on top of diapirs driven by salt movement at depth (Figs 9d and 15d).

The rate of sediment supply is also a key factor that dictates the pattern of mini-basin fill. Sediment supply is considered at the first order as a function of climate and tectonic forcing (Ethridge *et al.*, 1998). However, the occurrence of red floodplain mudstones with mud cracks and paleosols throughout the entire vertical succession of the Karayün Formation argues for a persistent semi-arid climate.

The ratio between generated accommodation space and sediment supply ( $A/S$ ) finds expression in the regional changes in the stacking patterns of sandstones bodies in the studied mini-basins (e.g. Martinsen *et al.*, 1999; Labourdette & Jones, 2007; Matthews *et al.*, 2007; Kjemperud *et al.*, 2008; Huerta *et al.*, 2011).

- A low sediment supply and/or a relatively high accommodation regime implies that the sediment supply cannot fill the available accommodation space. Under these conditions, the sedimentary sequence is typically composed of broad single-storey sheet-like sand bodies mainly isolated in a dominantly alluvial plain type setting containing mudstone, evaporites and lacustrine limestones such as observed in the playa-lake Lower member and saline lacustrine Upper member.
- In contrast, a high sediment supply and/or a relatively low accommodation regime implies that sediments fill the available space while eroding and bypassing the diapiric mini-basin borders. Under these conditions, the sedimentary sequence is typically composed of multi-storey sheet-like sand bodies that are vertically and laterally strongly amalgamated, containing a low proportion of preserved alluvial plain deposits such as found in the braided fluvial system of the Middle member.

Salt-induced local accommodation produced by salt evacuation can form topographically elevated salt walls or diapirs that impose a strong local control on sediment dispersal and the resulting facies distributions (Matthews *et al.*, 2007; Banham & Mountney, 2013a,b). The effects of topographically elevated salt walls and diapirs are especially well expressed when the sediment supply cannot fill the available accommodation space (Fig. 15a). In contrast, when the sediment supply is able to fill the available accommodation space induced by mini-basin subsidence, the salt walls and diapirs are buried and thus do not exert a local control on sediment dispersal (Fig. 15b).

The studied mini-basins formed in a foreland context and recorded the progression of deformation through time. The propagation of shortening in the Sivas Basin during the Oligocene is marked first by the abrupt transition from playa-lake dominated setting to a stacked fluvial braided system. This abrupt arrival of coarser sedimentation everywhere in the studied area is attributed to an

uplift in the hinterland of this Sivas foreland Basin. The progression of shortening is expressed by a rejuvenation of diapirs producing coeval large halokinetic folds from the end of the Middle member to the Upper member.

Three stages of development are proposed (Fig. 16a, c, d) for the Sivas mini-basins filled by the Karayün Fm. A missing intermediate stage is compiled by integrating several interpretations of analogous salt-controlled basins such as in the Paradox and La Popa Basins (Fig. 16b) to emphasize the influence of regional and local controls through time in the Sivas foreland Basin. Within these three basins, exclusively the ratio between the sediment supply and accommodation is compared to understand the various interactions with the salt tectonics in continental settings.

- During deposition of the playa-lake-dominated Lower member in the Sivas mini-basins, neighbouring mini-basins exhibited differing styles of infill with either sand-poor intervals characterized dominantly by fine-grained sheet-flood and evaporitic facies or sand-rich intervals with abundant conglomeratic lenses sandwiched between the floodplain facies (Figs 5 and 9b). The lower-slope and upper-slope units of the Chinle Formation in the northeast Paradox Basin also displays the same lateral evolution, with various facies assemblage variations between and within the mini-basins (USA; e.g. Matthews *et al.*, 2007). During periods of low sediment supply and/or relatively high salt-induced accommodation, the continental system is controlled by salt topography that governs the sediment dispersal pattern between paleotopographic diapir highs (Fig. 16a). This controlling factor is responsible for various facies changes and strongly modifies the architectural pattern from one mini-basin to another. The distribution of the facies assemblages is closely related to local depocentres, with preferential ponding of coarse sediments in the centre of the mini-basin (Fig. 16a). At the same time, the continued growth of the salt walls and the relative downbuilding of mini-basins result in the formation of successive small hook and wedge halokinetic geometries (i.e. tabular and tapered CHS) on the mini-basin flanks.
- Within the context of deformation migration from the south toward the mini-basins, it would be logical to assume an increase of the sediment-supply with relatively low salt-induced accommodation. This type of configuration is not observed in the Sivas mini-basins, but is recorded in the Black Ledge unit of the Chinle Formation (Matthews *et al.*, 2007), in the Triassic Moenkopi Formation (Banham & Mountney, 2013a) and in the proximal Permian Cutler Group (Cain & Mountney, 2009; Venus *et al.*, 2015) – all three occurring in the northeast Paradox Basin – and also in the Eocene Carroza Formation of the La Popa Basin (Andrie *et al.*, 2012). During these periods, the fluvial system progrades and expands across the overall area of the mini-basins (Fig. 16b). As salt-induced accom-

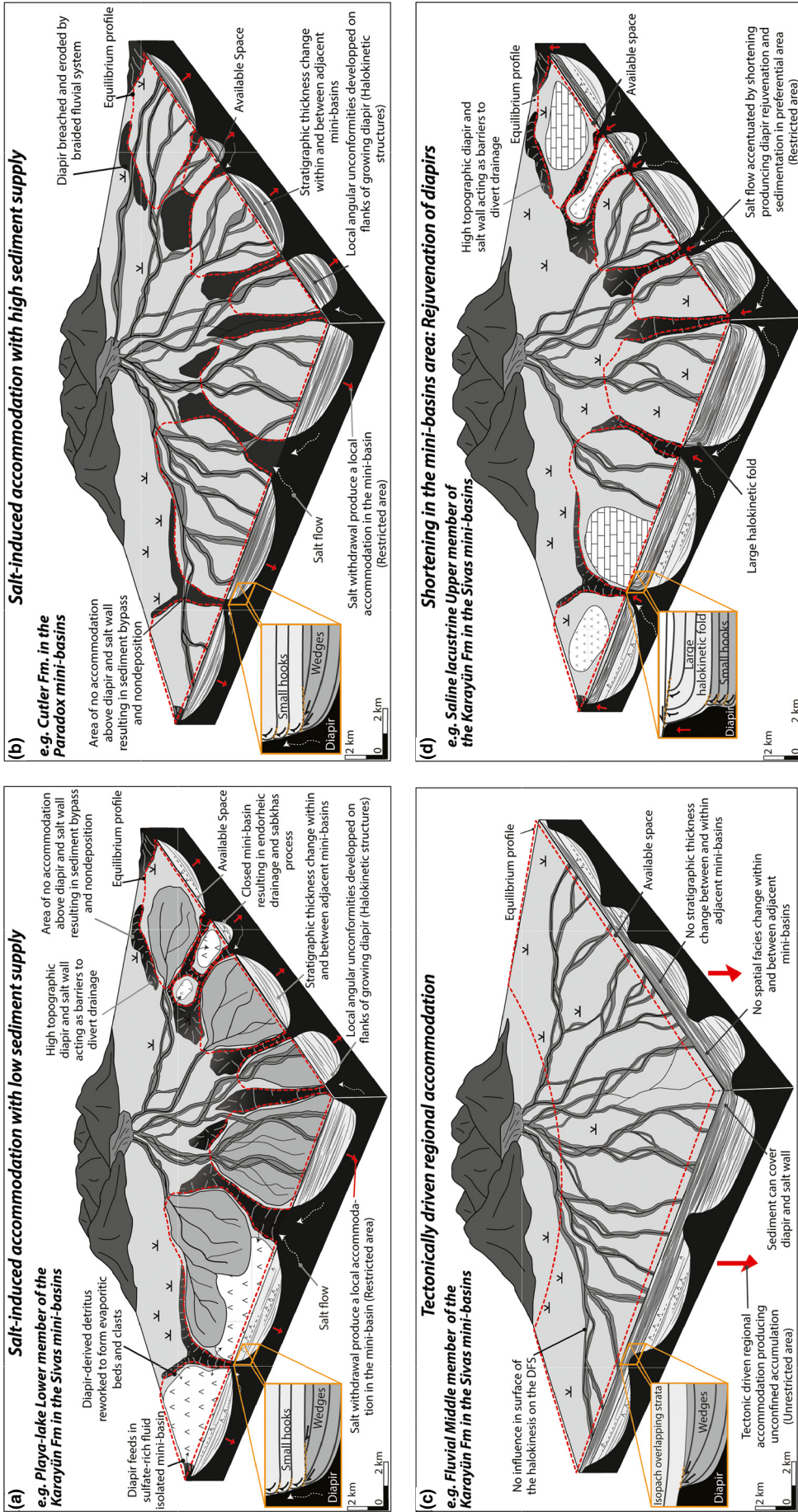


Fig. 16. Conceptual models of facies distribution, stratal geometries and thickness variations in a distributive fluvial system based on interpretations of the Sivas mini-basins as well as the Paradox and La Pops Basins. (a) Salt-induced local accommodation with low sediment supply leads to a strong influence of halokinesis on the paleotopography, by diverting the fluvial drainage and isolating certain mini-basins, as observed in the playa-lake Lower member. Strong changes in stratigraphic thickness also occur within and between the mini-basins, along with the development of halokinetic structures. (b) Salt-induced local accommodation with high sediment supply produces large changes in stratigraphic thickness owing to the influence of halokinesis, but has only a weak influence at the surface where salt structures correspond to areas of nil accommodation resulting in sediment by-pass (e.g. Matthews *et al.*, 2007; Banham & Mountney, 2013c). (c) Tectonically driven regional accommodation leads to a weak influence of halokinesis on the paleotopography, and sediments can cover the salt structures, such as observed during the initiation of the fluvial Middle member. (d) Shortening in the mini-basin domain produces diapir rejuvenation expressed by the occurrence of large halokinetic folds and topographic diapir highs, such as developed during deposition of the Upper member.

modation potential is directly linked to mini-basin subsidence, the salt walls and diapirs are considered as areas of zero accommodation with sediment bypass, and thus do not exert a significant paleotopographic control on sediment dispersal and distribution (Matthews *et al.*, 2007; Banham & Mountney, 2013c). Therefore, the mini-basins share common facies associations; however, distribution of facies assemblages between neighbouring mini-basins will not be influenced by salt-induced topography (Fig. 16b). Nevertheless, the distribution of the facies assemblage and the preserved thickness within a mini-basin depend on the differential creation of accommodation across the growing salt structures (Matthews *et al.*, 2007). At the same time, the continued growth of the salt walls and relative mini-basin downbuilding result in the formation of halokinetic structures on mini-basin flanks (Matthews *et al.*, 2007; Andrie *et al.*, 2012; Banham & Mountney, 2013a).

- During deposition of the Middle member, corresponding to a period of tectonically driven regional accommodation, the fluvial dominated braided depositional system accumulated vertically with little lateral variations in architecture or facies (Figs 9d and 16c). The resulting architecture is a relatively isopachous interval with little evidence of halokinetic stratal patterns such as small hooks and wedges. During this period of high regional accommodation rate and high sediment supply, a relatively isopachous interval of strata accumulated, draping and finally obscuring the salt topography (Fig. 16c). When accommodation rate and sediment supply are high, fluvial deposits tend to aggrade vertically and are widespread over the whole depositional area leading to a higher lateral continuity of sediment bodies between adjacent mini-basins, with limited effects of salt topography on drainage pattern (Fig. 16c).
- Strata at the top of the Middle member and within the Upper member are largely influenced by shortening in the mini-basin area, which reactivated salt flow in the diapirs. Large halokinetic folds are recorded on the flanks of the mini-basins (Fig. 16d). The rejuvenation of diaper-induced salt topography exerted a tight control on the sediment dispersal pattern, such as observed in the playa-lake facies of the Lower member. Neighbouring mini-basins exhibit differing styles of infill, with either sand-rich intervals characterized dominantly by fine-grained sheet-flood or sand-poor intervals with abundant evaporite and lacustrine carbonate beds (Figs 5 and 9d).

## CONCLUSION

- The depositional setting of the Karayün Formation is interpreted as a DFS with (i) the Lower member dominated by playa-lake and distal terminal splay deposits, (ii) the Middle member formed by highly

amalgamated fluvial braided deposits, and (iii) the Upper member consisting of saline lacustrine deposits. These three long-lived depositional systems fill ten mini-basins in the central part of the Sivas multi-basin domain and form large-scale stratal packages which are genetically related. The boundaries of these members are correlative basin-wide surfaces expressed by changes in texture and facies assemblages as well as stratal and stacking pattern.

- The variations in sediment supply and tectonically driven accommodation through time cause drastic changes in the architectural pattern of the Karayün Formation recorded coevally in all the mini-basins. The sharp transition between the Lower and the Middle member is characterized by a change in the channel type from isolated single-storey to multi-storey and multilateral sheet-like sandstones, with a net decrease in the proportion of floodplain deposits, interpreted as a basinward shift of facies (i.e. sequence boundary). The progressive transition from the Middle to the Upper member is characterized by a return to isolated single-storey units with a high preservation of floodplain mudstone deposits and the occurrence of evaporites and lacustrine carbonates.
- Tectonically driven regional accommodation (i.e. foreland basin subsidence) induces the deposition of relatively isopachous series, at the scale of the multi-basin domain, draping and finally obscuring the salt topography such as observed in the fluvial braided Middle member. The fluvial deposits tend to aggrade vertically, leading to a higher lateral continuity of sediments between adjacent mini-basins with salt topography having a limited effect on drainage pattern.
- When the accommodation is salt-induced, salt tectonics locally influences the stratigraphic thicknesses, the facies distribution and the stratal geometry at the mini-basin borders. Salt withdrawal below mini-basins results in differential local accommodation causing strong variations of stratigraphic thickness between adjacent mini-basins. Salt rise along the mini-basin edges results in halokinetic stratal geometries (i.e. small wedge and hook structures). When the sediment supply is low, transient topographies above salt ridges exert a local control on sediment dispersal. This paleotopographic control partitions and guides fluvial pathways across some mini-basins, while others become isolated from the drainage pathway, such as observed in the Lower and Upper members. On the contrary, when sediment supply is able to fill the available accommodation produced by mini-basin subsidence, the salt walls and diapirs show little topographic elevation and thus do not exert a local control on sediment dispersal. This is analogous to the situation observed in the Black Ledge unit of the Chinle Formation in the northeast Paradox Basin (Matthews *et al.*, 2007).



- Shortening in the mini-basin domain leads to a rejuvenation of diapirs and salt walls, thus producing large halokinetic folds observed in coeval deposits such as the Middle member.

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## CONFLICT OF INTEREST

No conflict of interest declared.

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