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Diagenesis of Oligocene continental sandstones in salt-walled mini-basins—Sivas Basin, Turkey

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ABSTRACT

The recent discovery of Oligo-Miocene salt-walled continental mini-basins in the Sivas Basin (central Anatolia, Turkey) provides the opportunity to unravel the influence of halokinesis on the diagenesis of continental mini-basin infilling. In this study, petrographic and geochemical analyses are used to define the diagenetic sequences recorded by two mini-basins filled mainly by fluvial clastic sediments of the upper Oligocene Karayün Formation. The initial diagenetic features are those commonly encountered in arid to semi-arid continental environments, *i.e.* clay infiltration, hematite precipitation and vadose calcite cement. Other early cements were strongly controlled by sandstone detrital composition in the presence of saline/alkaline pore water. In feldspathic litharenites and lithic arkoses, near-surface alterations were characterized by the precipitation of analcime (up to 10%), albite and quartz overgrowths (<1%). These events were followed by extensive calcite cementation (up to 30%) during shallow burial diagenesis which prevented further mesogenetic alteration phenomena such as compaction. In feldsarenites, early diagenesis differs by (i) the absence of analcime, (ii) better developed albite cements, (iii) thin smectite–illite coatings forming pore linings and (iv) patchy calcite cementation (<5%). The limited development of calcite cement allowed mesogenetic alterations to occur, such as late quartz overgrowths, albitization of feldspar grains and chemical compaction. All these phases are responsible for the low porosity of feldsarenites (<2%). The greater abundance of carbonate cement in feldspathic litharenites and lithic arkoses is related to a greater proportion of detrital limestone in these sandstones. Early precipitation of analcime, albite, smectite–illite and quartz was likely triggered by the alteration of reactive grains by near-surface saline/alkaline brines originating from the dissolution of adjacent diapiric structures. Mini-basin confinement resulting from halokinesis was probably an important factor influencing surface and subsurface saline/alkaline fluid flow and related diagenesis. Despite the evident role of detrital composition in controlling the recorded diagenetic evolution pathways, the striking similarity between observations in the Sivas Basin and equivalent halokinetic settings of the Pre-Caspian domain suggests that continental clastic deposits of mini-basins may exhibit common diagenetic alteration effects, especially near-surface zeolite precipitation. These effects are linked to the close proximity of diapiric salt during deposition and burial, and result in a significant degradation of porosity.

1. Introduction

Salt-withdrawal mini-basins form when marine to continental deposits subside into thick salt accumulations (Hudec and Jackson, 2007; Hudec et al., 2009). Salt expulsion and diapiric inflation produce complex stratal architectures that may act as efficient structural traps for potential hydrocarbon accumulations. Mini-basins have thus been increasingly targeted for oil and gas exploration in major hydrocarbon provinces such as the Gulf of Mexico, the west-African coast or Pre-Caspian Basin (Barde et al., 2002; Fort et al., 2004; Bartolini and Ramos, 2010).

In contrast to the large number of studies devoted to factors governing the generation and geometry of mini-basins (*e.g.* Rowan and Vendeville, 2006; Hudec and Jackson, 2009; Goteti et al., 2012; Warsitzka et al., 2013), only few studies have focused specifically on the diagenetic alterations affecting mini-basin clastic reservoirs (Underdown et al., 1990; Barde et al., 2002; Totten et al., 2007). Studies of salt dome provinces, such as the Gulf of Mexico, have shown that dissolution and dewatering of diapiric evaporites release large amounts of saline fluids (Ranganathan, 1992; Szalkowski and Hanor, 2003; Hanor and Bruno, 2014), whose migration into adjacent porous rocks is mainly driven by density contrasts with pore water (Ranganathan and Hanor, 1988; Warren, 2006). Fluid flow usually tends to focus along diapir flanks (i) because thermal anomalies surrounding salt domes increase pore-water density gradients, and (ii) because salt-sediment contacts may act as tectonic conduits (Evans and Nunn,

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1989; Reitz et al., 2007). Accordingly, the diagenetic alterations due to saline fluid migration have been studied mostly in sediments located in the immediate vicinity of salt domes (Leger, 1988; McManus and Hanor, 1993; Enos and Kyle, 2002; Totten et al., 2007). However, several studies have demonstrated that complex lateral kilometer-scale brine migrations may occur in permeable rocks (Bennett and Hanor, 1987; Bruno and Hanor, 2003; Hanor and Mercer, 2010). Such fluid flows may apply to mini-basins, but their impact on reservoir properties is poorly known (Underdown et al., 1990; Barde et al., 2002; Totten et al., 2007).

In continental environments, additional diagenetic complexity may result from the common dissolution of outcropping salt domes by meteoric water that leads to an increase in salinity of surrounding lakes and subsurface aquifers (Lawton and Buck, 2006; Warren, 2006, 2008; Buck et al., 2010; Mehdizadeh et al., 2015). This may induce (i) the formation of evaporitic paleosols, as observed in the La Popa and Paradox basins (Mexico and Utah, Buck et al., 2010; Lawton and Buck, 2006) or (ii) the precipitation of shallow burial zeolites, as reported in the Precaspian Basin (Barde et al., 2002). Thus, diagenesis of continental mini-basins may be affected by both surface and subsurface diapir-derived saline waters. The impact of such saline fluids on clastic reservoirs still needs to be clarified.

The Sivas Basin, located in the Central Anatolia Plateau of Turkey, contains some outstanding examples of outcropping salt tectonic structures with kilometer-scale mini-basins bordered by diapiric evaporites and filled with continental to marine deposits (Ringenbach et al., 2013; Callot et al., 2014; Ribes, 2015; Ribes et al., 2015a, 2015b; Kergaravat, 2016). The geometry, sedimentology and stratigraphy of upper Oligocene fluvio-lacustrine deposits filling several continental mini-basins have recently been constrained in detail in the central part of the Sivas Basin (Ribes, 2015; Ribes et al., 2015a, 2015b; Kergaravat, 2016).

The present study focuses on the diagenesis recorded by fluvial sandstones located in the main depocenter of two of these continental mini-basins (*i.e.* at some distance from the mini-basin borders). A combination of petrographic observations as well as geochemical analyses are used to (i) define the diagenetic sequences recorded in fluvial sandstones, (ii) constrain the conditions of precipitation of authigenic cements and (iii) evaluate the influence of the salt tectonic context on diagenesis.

2. Geological context

The east–west trending Sivas foreland Basin (~200 km by ~50 km) is located in the central part of the Anatolian Plateau in Turkey (Fig. 1) at the junction between three major crustal blocks, which comprise from north to south the Pontides thrust belt, the Kirşehir Massif and the Tauride–Anatolide block (Poisson et al., 1996; Görür et al., 1998; Yilmaz and Yilmaz, 2006; Rolland et al., 2010; Robertson et al., 2012, 2013). These blocks are separated by ophiolitic sutures (the Izmir–Ankara–Erzincan suture zone to the north and the Inner-Tauride suture zone to the south) resulting from the closure of the northern Neotethys Ocean during the late Cretaceous (Okay et al., 2006; Dilek and Sandvol, 2009; Rolland et al., 2010). The basement of the Sivas Basin is thus composed of metamorphic units of the Kirşehir Massif to the west (granitoid intrusions, marble, quartzite, gneiss, metapelitic schist and amphibolite; Lefebvre et al., 2011) and of obducted ophiolites to the east (Fig. 2).

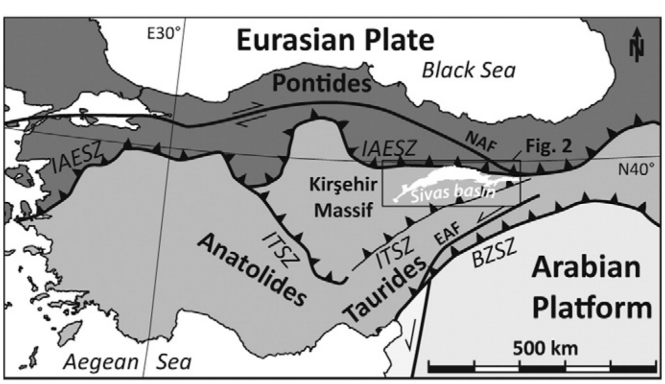

Fig. 1. Tectonic map of Eastern Mediterranean, with the main continental blocks (differentiated by grey tints), major suture zones and the Oligo-Miocene Sivas basin deposits (after Ribes et al., 2015a modified from Okay et al., 2006). BZSZ: Bitlis–Zagros Suture Zone, EAF: East Anatolian Fault Zone, IAES: Izmir–Ankara–Erzingan Suture Zone, ITSZ: Inner-Tauride Suture Zone, NAF: North Anatolian Fault.

Following obduction, the Sivas Basin evolved as a supra-ophiolitic flexural basin associated with a northward propagating fold and trust belt. From the upper Cretaceous (Maastrichtian) to the Eocene, early syn-tectonic deposits comprised (i) platforms fossiliferous to marly limestones along its southern highs (150–750 m, Gürlevik Formation), (ii) alluvial to deltaic conglomerates on the northern margin of the basin (500–1000 m, Bahçecik Formation), and (iii) deepwater volcanoclastics and turbidites in the east–west trending foredeep of the basin (1500–3000 m, Bözbel Formation) (Kurtman, 1973; Cater et al., 1991; Özçelik and Altunsoy, 1996; Poisson et al., 1996, 2010) (Fig. 3). Due to the northward propagation of the fold and thrust belt, the late Eocene records the uplifting of the basin and/or its closure from the marine domain. It resulted in a general shallowing of the area, marked by a transition from turbiditic to shallow marine environments, followed by the deposition of a thick evaporite succession emplaced in sebkhaic to salina depositional environments (Tuzishar Formation) (Poisson et al., 1996; Gündogan et al., 2005; Pichat et al., 2015). Due to the intense halokinesis and dissolution plus erosion processes (Gündogan et al., 2005; Kergaravat, 2016), the original thickness of the Tuzishar Formation is unknown. In the central part of the basin, it was thick enough to allow the onset mini-basin development during the Oligocene and the Miocene (Ringenbach et al., 2013; Callot et al., 2014; Kergaravat, 2016). After progressive continentalization of the basin, the northward to westward propagation of fluvial deposits (Selimiye Formation; 1500–2000 m) filled a first generation of mini-basins by sediment loading during the late Priabonian and Rupelian (Kergaravat, 2016). The top of the Selimiye Formation was marked by the emplacement of a large salt canopy overlying some of the first mini-basins (Ribes et al., 2015b; Kergaravat, 2016) (Fig. 3). During the late Oligocene, deposition of 1500 to 2000 m of fluvio-lacustrine sediments (Karayün Formation) associated to a distributary fluvial system was responsible for the development of a second generation of mini-basins above the canopy (Callot et al., 2014; Ribes et al., 2015b; Kergaravat, 2016). During the Early Miocene, following a regional marine transgression, the shallow marine Karacaören Formation (1000–2400 m) unconformably covered all previous formations and formed new mini-basins above the renewed salt canopy (Poisson et al., 2015; Ribes, 2015; Kergaravat, 2016). The marine deposits are composed of limestones, sandstones and marls, with small patch reefs of red algae and foraminifera (Kurtman, 1973; Cater et al., 1991; Poisson et al., 2015; Ribes, 2015). Following a last major regression, the marine facies were conformably covered by the Middle Miocene continental deposits of the Benilkaya Formation (500–1000 m) (Poisson et al., 2011; Ribes, 2015). Once again the facies characterize a distributive fluvial system overlain by continental evaporites and lacustrine carbonates (Poisson et al., 2010; Ribes, 2015) (Fig. 3).

This study focuses on two second-generation mini-basins, namely Emirhan to the West and Karayün to the East (Fig. 4). These basins are bordered by salt walls or welds. They were firstly filled by fluvio-lacustrine deposits of the mid–late Oligocene Karayün Formation, overlain by shallow marine deposits of the early Miocene Karacaören Formation. Emirhan and Karayün mini-basins are ~4 and ~5.8 km width respectively. The cumulative stratigraphic thickness of the fluvio-lacustrine deposits reaches ~2.4 and ~2.2 km in the Emirhan and Karayün mini-basins, respectively. The marine deposits account