



## Miocene facies associations and sedimentary evolution of the Southern Transylvanian Basin (Romania): Implications for hydrocarbon exploration

C. Krézsek<sup>a,\*</sup>, S. Filipescu<sup>b</sup>, L. Silye<sup>b</sup>, L. Mațenco<sup>a</sup>, H. Doust<sup>a</sup>

<sup>a</sup> Netherlands Center for Integrated Solid Earth Sciences, Vrije Universiteit, Faculty of Earth and Life Sciences, De Boelelaan 1085, 1081 HV Amsterdam, The Netherlands

<sup>b</sup> Babeș-Bolyai University, Department of Geology, Kogălniceanu 1, 400084 Cluj-Napoca, Romania

### ARTICLE INFO

#### Article history:

Received 8 September 2008

Received in revised form

14 July 2009

Accepted 30 July 2009

Available online 5 August 2009

#### Keywords:

Transylvanian Basin

Facies associations

Sequence stratigraphy

Hydrocarbon system

Biogenic gas

Exploration potential

### ABSTRACT

The Transylvanian Basin is a mature hydrocarbon province of Romania characterized by two petroleum systems: Mesozoic (thermogenic) and Miocene (biogenic). An extensive outcrop-based sedimentological and micropaleontological study correlated to seismic and well data discusses the elements of the Miocene petroleum system. The facies associations are indicative of alluvial, fandelta, shallow- and deep-marine settings. These are grouped into four different depositional systems (evaporite, mud-carbonate, sand-mud and sand-gravel). Their evolution in time and space shows large differences between various parts of the basin that have important consequences for exploration.

The Transylvanian gas is formed by more than 99% methane of bacterial origin. This is sourced by low quality (<1% TOC) deep-marine shales. The shales contain Type II and Type III kerogen. The organic material is thermally immature. The best source rocks were deposited during major transgressions in the central-eastern parts of the basin. In general, reservoir quality is the best (porosity < 20%, permeability < 1 D) in the basin center, where reservoirs are deep-marine turbidite sandstones. Lower quality reservoirs are conglomerate-rich slope channels and various shallow-marine sandstones located near the basin margins. The seals are formed by shales that hold gas columns of up to 60 m. The most common structural traps are in 4-way dip closures related to salt-cored folds. Their timing is coeval with the late (post-Pannonian) exhumation of the basin and strongly linked to coeval salt tectonics. This requires a late charge and migration.

The largest traps typically have multistory (up to 20) pay zones with a total of 100 BCF to 1 TCF reserves. Exploration to date has focused on structural traps, but most of the obvious structures have been drilled. It is argued that significant exploration potential lies in stratigraphic plays, including confined submarine fans, slope channels, detached lowstand prograding wedges, incised valleys, diapir flanks, salt-tectonics related unconformities and various sub-volcanic plays. Risks of the petroleum system elements associated to these plays in different areas of the basin are discussed.

© 2009 Elsevier Ltd. All rights reserved.

### 1. Introduction

The Transylvanian Basin (Fig. 1) is one of the most important gas provinces of Eastern Central Europe (Popescu, 1995). Exploration started 100 years ago and since then more than 90 gas fields have been discovered all located in the Miocene (Fig. 1; Ciupagea et al., 1970). However, only hydrocarbon shows have been encountered in the pre-Miocene deposits (Paraschiv, 1979).

The mid to late Miocene Transylvanian Basin is a semi-isolated back-arc basin, part of the intra-Carpathian back-arc system

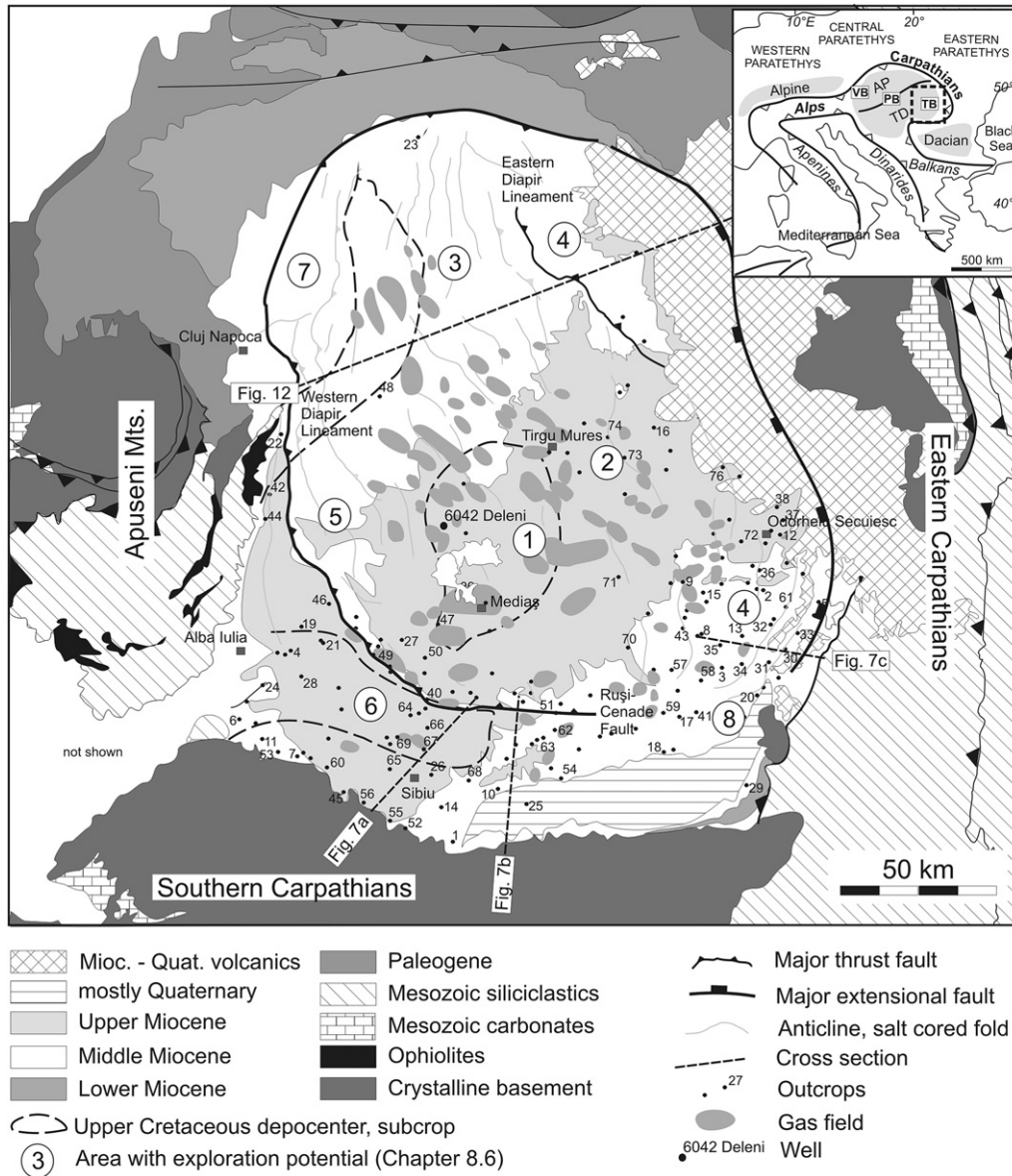
(see review in Schmid et al., 2008). The basin cannot be defined as intramontane, due to its large regional extent, nor open marine, due to its isolation reflected by the presence of endemic faunas (cf. Rögl, 1996). The basin has an area extent of ~200 × 250 km and was fed by short fluvial systems (~30 km) (Ciupagea et al., 1970). In an environment with gradually decreasing salinity (Ciupagea et al., 1970; Magyar et al., 1999), sedimentation was essentially aggradational and major progradation occurred only during relative sea-level falls (Krézsek and Filipescu, 2005).

Exploration in Transylvania typically consisted of drilling structural closures (e.g. 4-way dip closure; Paraschiv, 1979). At present, however it is becoming increasingly difficult to find significant structural traps as shown by the exploration results of the last 30 years (Nemeșan, 2007).

In this paper, our scope is to show evidences of under-explored and new play types in this mature basin. One major impediment is

\* Correspondence to: Present address: Csaba Krézsek, Petrom S.A., E&P Headquarters, Eroilor Square 1A, RO-100316, Ploiesti, Romania. Tel.: +40 372 484774; fax: +40 244 549702.

E-mail address: [csaba.krezsek@dal.ca](mailto:csaba.krezsek@dal.ca) (C. Krézsek).



**Fig. 1.** The Transylvanian Basin (TB) is the easternmost Intra-Carpathian basin (see inlet) developed on top of the Tisza–Dacia block (TD). This latter and the Alcapa (AP) are two major tectonic units assembled by the Mid-Cretaceous and form the basement of the intra-Carpathians basins as the Pannonian Basin (PB) and Vienna Basin (VB) (e.g. Csontos and Vörös, 2004; Schmid et al., 2008). The Transylvanian Basin has a roughly circular shape surrounded by the Apuseni Mts., Eastern and Southern Carpathians. The simplified geological map illustrates the main N/NW to S/SE fold trends of the basin characteristic for the gas-bearing Middle to Late Miocene deposits. The pattern of gas field outlines follows the above structural trend. Most important Miocene structural features are highlighted including the Western and Eastern Diapir Lineaments, and the Ruși-Cenade Fault. The encircled numbers indicate play trends and associated risks in the Transylvanian Basin based on the composite source, reservoir, seal and trap risk assessments (see Chapter 8). The central part of the basin (Area 1) is the most prolific as shown by the number of gas fields. However, due to the intensive exploration further potential is limited to small closures and stratigraphic pinch-outs of submarine fans. In Area 2 and Area 3 untested play type is the upslope pinch-out of slope fans and channels. The eastern part of the basin (Area 4) has a variety of interesting, yet untested plays including sub-volcanic structural closures and stratigraphic traps on diapir flanks. In Area 5 thin-bedded turbidite reservoirs and detached shallow-marine offshore bars may be exploration targets. Area 6 may hold a reasonable exploration potential in small structural closures. Limited exploration potential characterizes areas 7 and 8. Annotations: WDL – Western Diapir Lineament, EDL – Eastern Diapir Lineament, RC – Ruși-Cenade Fault. The map also shows the numbered location of most important outcrops presented or discussed here: 1 Tâlmaci, 2 Ulieș, 3 Dacia, 4 Berghin, 5 Crăciunel, 6 Petrești, 7 Dobârca, 8 Bunești and Mihai Viteaz, 9 Mureni, 10 Săcădate, 11 Călnic and Rahău, 12 Brădești and Satu Mare, 13 Beia, 14 Mohu, 15 Archita, 16 Ghindari, 17 Felmer, 18 Făgăraș, 19 Colibi, 20 Ionești, 21 Tău, 22 Săndulești, 23 Nireș, 24 Daia Română, 25 Sărata, 26 Gușterița, 27 Micăsasa, 28 Vingard, 29 Perșani, 30 Mercheaș, 31 Homorod, 32 Comana, 33 Jimbor, 34 Rupea, 35 Fișer, 36 Nicolești, 37 Săncraia and Tărnovița, 38 Dealu, 39 Bazna, 40 Ruși, 41 Cobor, 42 Gârbova de Jos and Gârbovița, 43 Criț, 44 Lopadea Veche, 45 Sibiel, 46 Mihălț, 47 Coșpa Mică, 48 Hădăreni, 49 Cenade, 50 Agărbiciu, 51 Vărd and Bârgăniș, 52 Călnădioara, 53 Rod, 54 Arpașu de Jos, 55 Rășinari, 56 Fântânele, 57 Grnari, 58 Jibert, 59 Rodbav, 60 Apoldu de Sus, 61 Ungra, 62 Chirpăr, 63 Ilimbav and Marpod, 64 Soroștin, 65 Ocna Sibiu, 66 Slimnic, 67 Șura Mare, 68 Cornățel, 69 Loameș, 70 Brădeni, 71 Sighișoara, 72 Feliceni and Mugeni, 74 Solocma, 75 Sântana de Mureș, 76 Praid.

that the elements of the petroleum system are poorly understood despite more than 4000 wells have been drilled (Paraschiv, 1979). Therefore, our approach is to describe first the facies associations in the basin based on an extensive outcrop study. This enables us to discuss the distribution, thickness and quality of petroleum system elements in the basin, which ultimately provides background for play types and exploration potential of the Transylvanian Basin.

## 2. Overview of the regional evolution

The Carpathian orogen is part of an Alpine system that resulted from the Triassic to Cenozoic evolution of continental blocks derived mostly from the European margin (Săndulescu, 1988; Schmid et al., 2008). The middle-late Miocene Transylvanian Basin overlaps an earlier Cretaceous orogenic evolution, which led to the closure of the Transylvanides (or East Vardar) oceanic domain and convergence between Tisza and Dacia intra-Carpathian blocks (Fig. 1; Săndulescu, 1988; Cloetingh et al., 2004; Schmid et al., 2008). The subsequent Middle to Late Miocene evolution is linked to the gradual movement of the upper Carpathian units into the so called Carpathians embayment, a 160–220 km wide continental margin (e.g. Roure et al., 1993) adjacent to the European foreland. The movement of the Carpathian units was enhanced during the Miocene by slab roll-back in the distal parts of the European foreland (e.g. Royden, 1988; Wortel and Spakman, 2000). As a consequence, shortening in the East Carpathians during the Middle Miocene was coeval in its hinterland with the extensional collapse of the Pannonian Basin (e.g. Horváth 1993; Tari and Horváth, 2006) and regional subsidence which led to the deposition of more than 3 km of siliciclastic sediments in the Transylvanian Basin (Fig. 2a). The onset of Carpathians collision with the East-European craton and its southern platform units at the end of the Middle Miocene and beginning of the Late Miocene led to ~5 km exhumation in the core of the central East Carpathians (Sanders et al., 1999; Gröger, 2006). This was followed by late Miocene exhumation/erosion of Transylvania to the present-day elevations of more than 400 m. Miocene basin evolution and its subsequent inversion was roughly coeval with onset of back-arc calc-alkaline volcanism (Pécskay et al., 1995; Szakács and Seghedi, 1995; Mason et al., 1998).

The uplift of the Alps and Carpathians severely limited marine connections with the open ocean since the beginning of the Oligocene and triggered restricted conditions and endemic evolution of fauna north of the Mediterranean realm (e.g. Rögl, 1996), where the Paratethys intra-continental sea developed (Steininger and Senes, 1971). During its Miocene evolution, Transylvania was part of the Central Paratethys, which also incorporated the larger Pannonian Basin (Fig. 2a, Popescu et al., 1995; Harzhauser and Piller, 2007).

During the Early Miocene continental settings prevailed in the basin, with the exception of its northern part, where a foredeep wedge continued to evolve as a result of continental collision between the ALCAPA and Tisza–Dacia blocks (Fig. 1; Tischler et al., 2008). The overall continental conditions changed significantly by the early Mid Miocene (Early Badenian), when a regional transgression established relatively deep-marine conditions in the basin. Mixed carbonate – siliciclastic platforms formed on the basin margins and siliciclastic sediments were deposited in deeper areas. The overall subsidence was rather low, and accommodated up to 100–200 m of sediments in the basin (Krézsek and Filipescu, 2005).

Highly restricted conditions developed during the Mid Miocene (Peryt, 2006), which led to evaporite deposits, reaching 300 m in the Transylvanian Basin, regionally distributed with the exception of a few basement highs. The evaporites consist of gypsum on the basin margins, and halite interbedded with thin shales in the basin center, the entire package being generally referred as the “Middle

Badenian salt” (Ciupagea et al., 1970). The salt had a very important role in facilitating deformation by detaching the post-salt sedimentary package from the pre-salt basement as a gravitational slide that affected the whole basin during the Late Miocene (Krézsek and Bally, 2006).

Post-salt subsidence and depositional rates dramatically increased in the basin during the upper part of Middle Miocene and the lower part of Late Miocene, i.e. Upper Badenian to Pannonian (e.g. Crănganu and Deming, 1996). Although extensional phases have been speculated in particular for the Middle Miocene (e.g. Huismans et al., 1997; Huismans and Bertotti, 2002), modern subsurface imaging have revealed the absence of significant Miocene normal faults in the entire basin that could account for the observed subsidence rates (Krézsek and Filipescu, 2005). The subsidence wavelengths are in the order of 200–300 km and accordingly suggest sub-crustal mechanisms (Cloetingh et al., 2006; Maţenco et al., 2007). In this context, a range of sub-lithospheric processes such as thermal doming related to simple-shear extensional collapse of the Pannonian basin (see Tari et al., 1999 for the kinematics of the latter) or thermal anomalies linked to the gradual Miocene evolution of the presently observed Vrancea slab (Ismail-Zadeh et al., 2005) can be discussed.

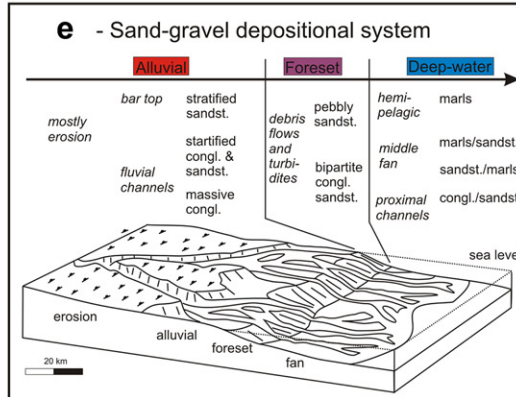
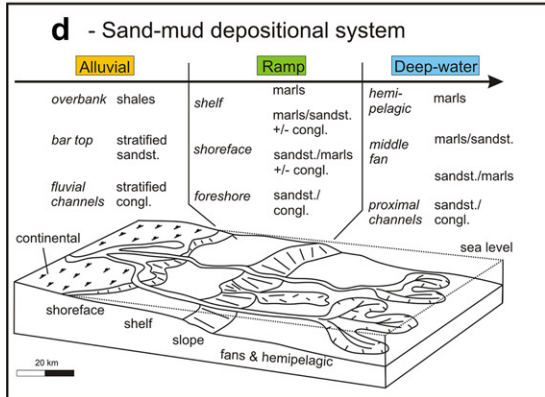
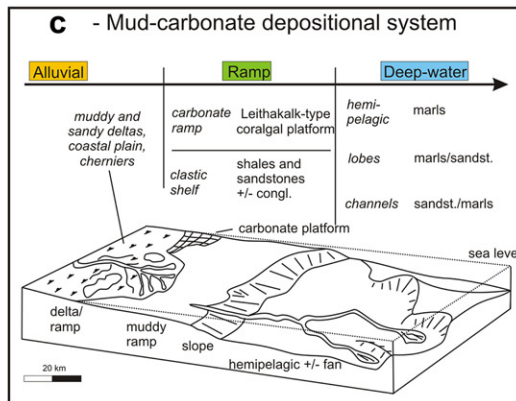
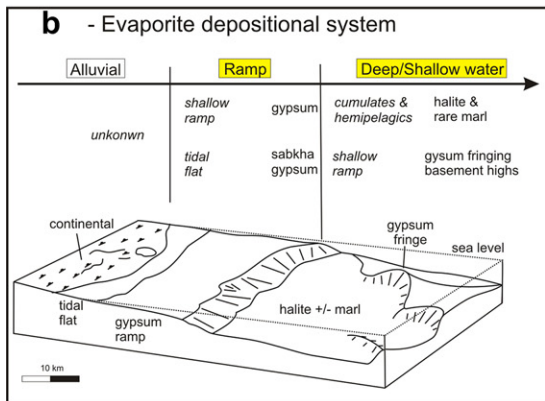
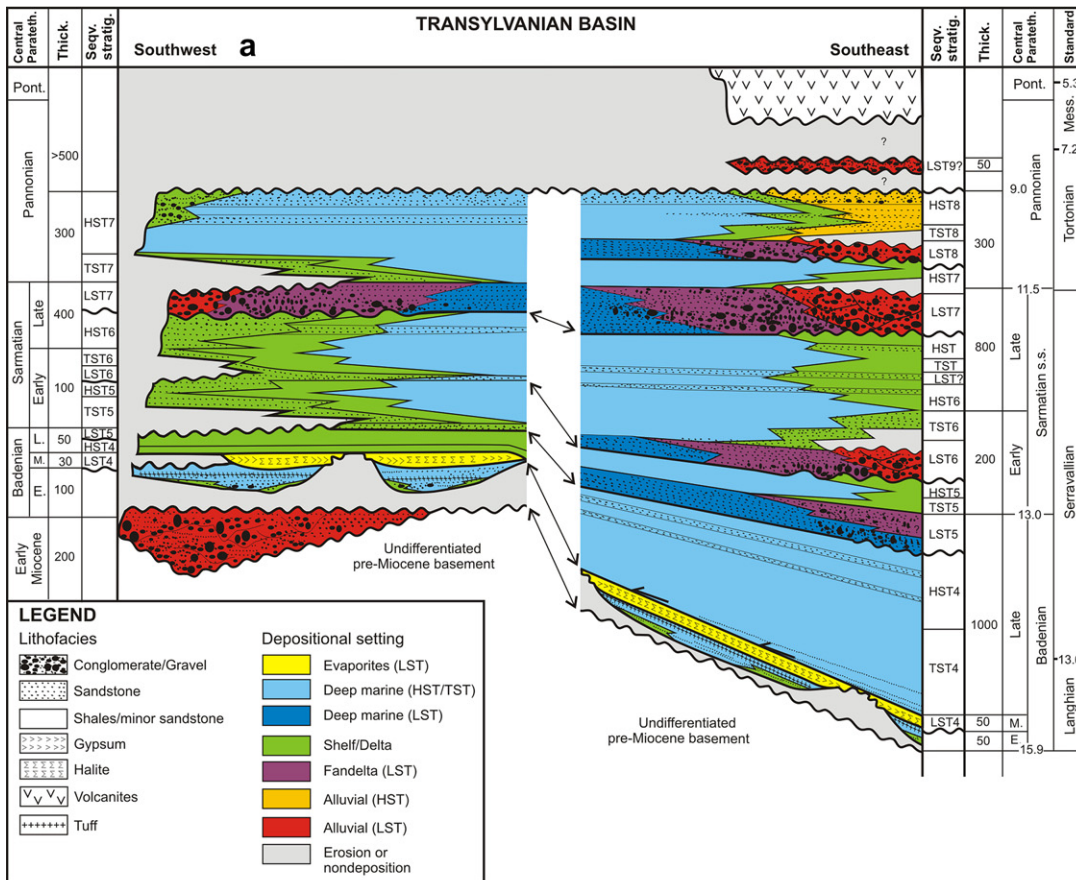
The increase of post-salt depositional rates was coeval with the shortening in the external, thin-skinned part of Carpathians and the onset of exhumation in its orogenic core. Up to 6 km of sediments were eroded in the central part of the East Carpathians during the Middle–Late Miocene (Sanders et al., 1999). As a result, the margins of the basin, with their inner Upper Badenian–Sarmatian marine/deltaic deposits, have been severely affected by later exhumation and erosion.

Late Mid Miocene (end of Sarmatian) uplift of the Carpathians, due to continental collision, completely cut off the intra-Carpathian realm from the rest of the Paratethys. As a consequence, the Transylvanian Basin became an embayment of the intra-Carpathian Lake Pannon (Magyar et al., 1999). Final infill of the basin did not take place until the end of the Miocene, but these deposits have been hardly preserved except where they are covered by back-arc volcanics (Fig. 1; Krézsek and Filipescu, 2005).

## 3. Data and methods

Our interpretation relies on an extensive outcrop-based study. More than 200 outcrops have been studied mainly in southern part of the Transylvanian Basin (Fig. 1). The sedimentological descriptions led to the identification of 20 types of facies that comprise various facies associations indicative of broad depositional settings, from alluvial to deep marine (Table 1). Wherever the clastic material and its content was suitable, samples were analyzed for biostratigraphic and paleoenvironmental information. Micropaleontological data diagnostic either for marine (e.g. foraminifera) or restricted brackish to lacustrine (e.g. ostracodes) environments have been used for biostratigraphic correlations and to constrain the described depositional environments.

The outcrop study was focused near the mountain borders to detect marginal unconformity surfaces and major flooding surfaces. These were correlated to distal basinal areas using seismic lines and well logs, where the unconformities laterally change into correlative conformities. This analysis has been integrated into the sequence stratigraphic framework of Krézsek and Filipescu (2005) where transgressive surfaces and maximum flooding zones were correlated in well logs across the basin to derive basin evolution. This is because flooding surfaces were easier to correlate basin-wide than the sequence boundaries. Many sequence boundaries are only locally developed and reflect tectonic events in the Carpathians (Krézsek and Filipescu, 2005).



Our sedimentological study was integrated into the regional tectonic and sequence stratigraphic framework of the basin (Krézsek and Filipescu, 2005). This details the depositional systems and develops depositional models to reconstruct the paleogeography (Section 6). Then these reconstructions complemented with other data, were used to discuss the Miocene petroleum system (Sections 7 and 8).

#### 4. Facies associations

The sediments studied in the southern part of the Transylvanian Basin are conglomerates, sandstones, shales, gypsum and halite. These form characteristic sedimentary successions exposed in several outcrops (Fig. 1), which are grouped into the facies associations described below.

##### 4.1. Facies association I

Consists of reddish conglomerates with minor amount of pebbly sandstones and sandstones (Fig. 3a–d), and rare shales (e.g. Sâncrai, Târnovița, Tâlmaci; Fig. 1). The conglomerates are pebble- to boulder-sized, dominantly clast-supported reddish assemblages, massive or stratified, and form several meter thick beds. The massive conglomerates display lower erosional boundaries and are fining upwards into pebbly sandstones. The vast majority of the clasts are imbricated parallel to their long axis [a(p)a(i)]. The stratified conglomerates exhibit well or vaguely defined horizontal-, planar- or trough-stratification (Fig. 3a–c). These beds have sharp lower and upper boundaries.

Conglomerates are typically overlain by thinner units of pebbly sandstones, sandstones or rarely thin shales (Fig. 3a). The sandstones display horizontal-, planar- or trough-stratification and frequently contain discontinuous rows or scattered pebble- to cobble-sized lags (Fig. 3d). The boundary between the sandstones/shales and conglomerates is in general sharp and erosional.

##### 4.1.1. Interpretation

The massive and clast-supported conglomerates, which lack any stratification, suggest deposition from hyperconcentrated flows as indicated by the pebble's imbrication. Stratification in conglomerates and sandstones is the effect of unidirectional flows and may be interpreted as fluvial deposits. Individual fluvial bars are built by a lower unit of stratified conglomerates (gravelly bar) and an upper unit of stratified sandstones (sandy bar top formed by fluvial dunes) occasionally covered by shales. These shales are interpreted as deposited in abandoned channels or as overbank fines.

The conglomerates, sandstones and shales form up to 10 m thick fining-upwards successions interpreted as alluvial channel fills. The

products of hyperconcentrated flows are situated in the lower part of the channels. This suggests, the channels were initiated by the hyperconcentrated flows and subsequently filled by stacked gravelly and sandy bars. In general, the mass transport deposits are more frequent than the gravel bars, but significant variability is observed in individual successions.

The observed facies association is comparable with high-energy depositional environments diagnostic for gravel-rich braided rivers (Fig. 3b, c; Blair and McPherson, 1994). The dominance of the mass transport deposits to traction deposits would indicate the presence of the source nearby, i.e. deposition on proximal parts of alluvial fans. The continental setting is supported by the lack of any types of marine fauna and the overall reddish color of the sediments.

##### 4.2. Facies association II

This is composed of conglomerates, gravel lags, pebbly sandstones, sandstones, some of them highly bioclastic, and significant amount of siltstones, marls and shales (e.g. Dobârca, Petrești, Crăciunel; Figs. 1, 4a–j). These form characteristic assemblages, as follows:

- Horizontally laminated and coarse-grained sandstones with scattered pebble lags and some minor lenticular conglomerates (Fig. 4a, b). The pebble lags show [a(t)b(i)]-type imbrication (e.g. Crăciunel). The conglomerates frequently grade upwards into pebbly sandstones and sandstones.
- Sandstones, sometimes pebbly, dominated by up to 1 m thick planar cross-stratification, associated with scattered discontinuous lags represented by reworked imbricated mud chips or gravel pebbles (Fig. 4c). Low-angle erosional surfaces are very frequent associated with pebbly sandstones (Petrești, Fișer) or even conglomerates (Dobârca).
- Well-sorted and bioclastic, up to a few decimeters thick sandstones with thin bioturbated shale caps (Fig. 4d). Bidirectional cross-lamination and hummocky stratification are widespread (Fig. 4d–f). The shales may be deeply eroded and preserved only as discontinuous patches between sandstones or may form a significant part of the facies association (Dobârca, Sărata).
- Massive or laminated shales and thin unidirectional cross-laminated sandstones, which alternate sometimes with a minor amount of bioclastic sandstones (Fig. 4g, h), pebbly sandstones (Fig. 4i) or even microconglomerates (Fig. 4j). The marls tend to be more carbonate rich with millimeter-thick white carbonate laminae giving an outcrop white-grey striped pattern (e.g. Dobârca). The coarse-grained sandstones have frequent bioclastic constituents and are formed by a lower horizontally

**Fig. 2.** Sedimentary evolution and depositional models of the Southern Transylvanian Basin. (a) Sketch of the depositional settings and sequence stratigraphy. The systems tract annotation follows Krézsek and Filipescu (2005). It also shows the correlation of the Central Paratethys vs. Standard chronostratigraphy (Rögl, 1996; Gradstein et al., 2004). The color code matches the depositional settings of Fig. 2b–e. It highlights the main differences between the southwestern and southeastern parts of the basin based on this study. The sedimentary succession in the east is thicker and more sand-prone than to the west. Note the thick Upper Badenian and the well developed and conglomerate-rich lowstand systems tracts in the eastern part of the basin. Note also, that the outcropping Pannonian mostly consists of deep-lacustrine deposits. The final shallow-marine to alluvial fill of the basin has not been preserved except to the east, where a younger volcanic cover prevented erosion. (b) The evaporite depositional system is formed by tidal flats, shallow platforms and deep-marine ramps. Sabkha-type gypsum has been deposited on narrow tidal flats that fringed wider, but less than 10 km shallow-marine gypsum platforms. These deposits have been preserved in the western part of the basin. The central parts of the basin were dominated by deep-marine halite cumulates. A minor amount of gypsum likely was formed around some exposed basement highs. (c) The mud-carbonate depositional system is assembled by shallow-marine carbonate or mud-rich ramps and mud-rich submarine fans. This depositional system represents the Early Badenian. The carbonate platforms were formed by corallgal buildups and are locally found in the western part of the basin. Elsewhere muddy ramps developed with small deltas. The deltas fed mud-rich submarine fans in the central parts of the basin. (d) The sand-mud depositional system is the predominant depositional system of the basin that prevailed during sea/lake-level highstands. It is formed by a wide range of facies associations deposited in alluvial, shelf and deep-marine/lacustrine environments. Note that the alluvial part has been only identified in the Pannonian. See the text for details. (e) The sand-gravel depositional system characterizes sea/lake-level lowstands. It is formed by braided alluvial systems, fandelas and sand-rich submarine fans. Some braided alluvial systems have been deposited in incised valleys cut into the previous highstand coastal plain/shelf (e.g. LST6–8 in Fig. 2a). Characteristic elements of this depositional system are the gravelly or sandy shelf-edge fandelas best developed during LST7 and 8 (Fig. 2a). The fandelas supplied coarse-grained sediments to the sand-rich submarine fans in the basin center.

**Table 1**  
Lithofacies associations, description and interpretation of facies associations, microfauna and representative outcrops of the Southern Transylvanian Basin. The location of the outcrops is shown in Fig. 1. Annotations: LM2 – upper part of the Lower Miocene, Ba1 – Lower Badenian, Ba2 – Upper Badenian, Sa1 – Lower Sarmatian, Sa2 – Upper Sarmatian, Pa – Pannonian.

Lithofacies	Description	Interpretation	Microfauna	Representative outcrop and their age
<b>Fluvial</b>				
Massive conglomerate	Meter-scale granule to boulder-size clast or matrix supported conglomerates. Ungraded/graded $\pm$ crude horizontal imbrication, channelised or irregular base.	Channel fill debris flows and longitudinal bedforms.	Absent	Ba1/LM2?: Tălmăciu Sa1: Dacia, Beia Sa2: Ulieș
Startified conglomerate	Meter-scale pebble- to cobble-sized clast-supported conglomerates. Graded with planar cross-stratification or through cross-stratification. Frequently scoured base.	Transverse bars or minor channel fills.	Absent	Pa: Satu Mare, Târnovița, Sâncrai, Dealu
Gravel lag	1–2 clast thick row of imbricated pebbles and cobbles that form discontinuous patches at the base of stratified sandstones or conglomerates.	Minor channel lag.	Absent	Frequent
Stratified sandstone	Decimeter-scale medium to coarse-grained, sandstone and pebbly sandstone. Fining upwards, planar and through cross-stratification, horizontal lamination.	Sand dunes (bar top assemblages), low-energy channel fill.	Absent	Frequent
Shale and minor fine sandstone	Decimeter-scale massive brown claystone and minor siltstone, interbedded with minor horizontally or ripple-laminated fine-grained sandstone and discontinuous coal seams.	Overbank fines, channel abandonment.	Absent	Pa: Târnovița, Sâncrai
<b>Shallow-marine ramp</b>				
Conglomerate	Decimeter- to meter-scale pebble- to boulder-sized clast-supported conglomerates with a coarse sandy to pebbly sandstone matrix; erosional lower boundaries and highly-variable thicknesses; vaguely defined tabular or horizontal stratification; some may be massive and fining upwards.	Distributary channels incised on the inner shelf (mostly upper shoreface).	Absent	Sa2: Berghin, Crăciunel, Rahău Pa: Vingard
Pebbly sandstones	Decimeter-scale well-sorted coarse-grained horizontally stratified sandstones with scattered granule to boulder-sized pebble lags. The stratification is formed by centimeter-scale fining- or coarsening-upwards units. Large-scale low-angle cross-stratification and low-angle erosional surfaces.	Foreshore/Upper shoreface pebbly sandstones. Frequently associated with major conglomerate channels and gravel lags.	Absent	Sa2: Crăciunel, Berghin, Dobârca valley Pa: Vingard
Coarse sandstone with gravel lag	Decimeter-scale beds up to 1–2 m thick of medium to coarse-grained, well-sorted and mostly fining-upwards sandstones; occasionally pebbly and with minor lenticular conglomerates. Tabular and swaley cross-stratification, minor horizontal lamination. Low-angle truncations and gravel lags are frequent.	Large-scale upper shoreface dunes with gravel lags and minor conglomerate channels.	Absent	Sa1: Daia Română, Petrești; Fișer, Bunești, Dobârca valley Sa2: Crăciunel, Cornățel, Mureni Pa: Vingard
Bioclastic sandstone and shale	Centimeter to decimeter-scale coarse to medium grained, frequently bioclastic sandstones alternating with centimeters thick shale; the sandstone/shale boundary is sharp; rare pebble lags; the upper surface of the sandstones is not graded; wave-current ripples, hummocky cross-stratification, horizontal lamination; shales are bioturbated.	Minor lower shoreface sandy dunes with suspension fall-out marls.	Sa1: shallow-water benthic foraminifera (e.g. <i>Ammonia</i> spp., <i>Elphidium</i> spp., <i>Nonion</i> spp., miliolids), mysids, and ostracods Sa2: shallow-water benthic foraminifera (e.g. <i>Ammonia</i> spp., <i>Porosonion</i> spp.)	Sa1: Sărata, Dobârca valley Sa2: Mureni
Marl with sandstone and conglomerate	Centimeters-scale laminated brown and whitish marls/shales and ripple-laminated fine-grained sandstones that form several meter thick successions occasionally interbedded with centimeter-scale fining-upwards bioclastic sandstones and conglomerates; sandstones have horizontal to ripple lamination; conglomerates are massive or planar cross-stratified.	Inner shelf below the storm-wave base with frequent tempestites.	Shallow-water microfauna Sa1: e.g. <i>Porosonion</i> spp., <i>Elphidium</i> spp.), ostracods Sa2: e.g. <i>Porosonion</i> spp., <i>Nonion</i> spp., <i>Bolivina</i> spp., ostracods	Sa1: Bunești, Dobârca valley Sa2: Mureni, Ulieș, Dobârca valley Pa: Gârbova de Jos, Gârbovița, Lopadea Veche 1, Săcădate
Marls with fine-grained sandstone	Centimeter-scale marls locally in alternation with minor fine-grained sandstone; unidirectional ripples and horizontal lamination, carbonate rich laminae.	Inner to outer shelf below the storm-wave base.	Sa1: benthic foraminifera (e.g. <i>Elphidium</i> spp., <i>Nonion</i> spp., <i>Cassidulina</i> spp., <i>Bolivina</i> spp., <i>Anomalinoidea dividens</i> , miliolids), rare planktonic foraminifera (e.g. <i>Tenuitellinata</i> spp., <i>Globigerina</i> spp.), ostracods, and juvenile mollusks Sa2: mysids, ostracods	Ba2: Rod Sa1: Dobârca valley, Arpașu de Jos, Rășinari, Fântânele, Mihai Viteazu, Grânari, Apoldu de Sus Sa2: Mureni 2
<b>Fandelta</b>				
Bipartite conglomerate and sandstone	Decimeter- to meters-scale fining-upwards, clast or matrix supported pebble- to boulder-sized massive conglomerates, with [a(p)a(i)]-type imbrication that form multi-storey amalgamated units. The conglomerates are covered by decimeter thick horizontally laminated or massive coarse-grained sandstones. Pebble- to boulder-sized imbricated lags are frequent; high-angle depositional dips.	Bipartite flows consisting of lower cohesionless debris flows and upper turbidity currents and debris falls deposited on gravelly Gilbert-type fandelta foresets	Absent	Sa2: Nicolești, Ulieș, Lopadea Veche Pa: Brădești, Mugeni, Feliceni

Table 1 (continued)

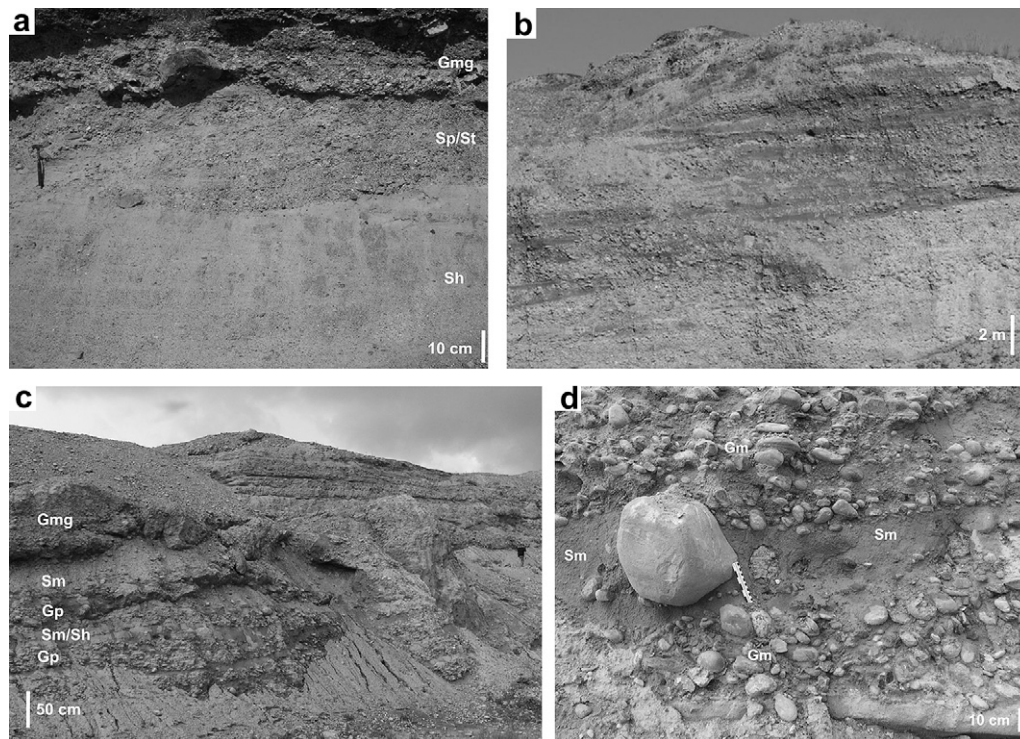
Lithofacies	Description	Interpretation	Microfauna	Representative outcrop and their age
Pebbly sandstone and sandstone	Decimeter- to meters-scale massive or horizontally laminated pebbly sandstones that fine upwards into sandstones. Frequent imbricated [a(p)a(i)] pebble lags and shale rip-ups; rare lenticular cross-stratified pebble conglomerates; low-angle depositional dips and erosional surfaces; bioturbation.	High-density turbidites or sandy debris flows and debris falls probably deposited on sandy fandelta low-angle foresets. Minor conglomerate channels.	Absent	Sa2: Mohu, Călnic
<b>Deep-water</b> Massive sandstone and conglomerate	Several meters thick fining-upwards massive coarse-grained sandstones with frequent internal amalgamation surfaces sometimes capped by low-density turbidites (Tbc). Frequent shale rip-ups. The sandstones may be interbedded with meters thick channelised units formed by pebble- to boulder-sized, clast or matrix supported, frequently imbricated [a(p)a(i)] conglomerates.	Proximal submarine channel fills formed by high-density turbidites, gravely debris flows and some low-density turbidites.	Absent	Ba2: Jimbor Sa1: Cobor Sa2: Archita Pa: Ghindari
Sandstones with minor marls	Decimeter to several meter-scale Bouma-type (Tab, Tabc, Tbc) sandy turbidites interbedded with decimeter thick marls. Non-erosional parallel bed boundaries. Rare small-scale sandy channels and shale rip-ups.	Submarine channels late fill or submarine fan lobe dominated by low-density turbidites and interbedded with suspension fall-out pelagics. High-density turbidites may occur.	Ba2&Sa1: planktonic foraminifera ( <i>Tenuitella</i> spp., <i>Tenuitellina</i> spp., <i>Globigerina</i> spp.), rare benthic foraminifera (e.g. <i>Bolivina</i> spp., <i>Oolina</i> spp., allochthonous juvenile specimens) Sa2: rotallids, and ostracods	Ba2: Comana, Ungra, Homorod, Rupea, Sa1: Felmer, Făgăraş Sa2: Rodbav, Bărghiş, Cenade, Chripăr, Criş, Ilimbav, Jibert, Marpod Pa: Tău, Sântana de Mureş, Solocma, Loamneş, Ocna Sibiu, Ruşi, Copşa Mică, Micăsasa, Sighişoara, Slimnic, Şura Mare, Feliceni
Marls with minor sandstones	Decimeter to meter scale laminated or massive marls associated with minor fine to medium grained Bouma-type (Tbc, Tcd) sandy turbidites. Non-erosional parallel bed boundaries. No rip-ups or channels.	Submarine fan levee or outer lobe fringe; dominated by suspension fall-out with minor amount of low-density turbidites.	Sa1: planktonic foraminifera, benthic foraminifera (e.g. <i>Bolivina</i> spp., <i>Oolina</i> spp., <i>Anomalinoidea dividens</i> , ?rare agglutinated, allochthonous juvenile specimens of rotallids?), ostracods Sa2: benthic foraminifera (e.g. <i>Ammonia</i> spp., <i>Elphidium</i> spp., <i>Porosonion</i> spp.), and ostracods Pa: ostracods	Ba2: Mercheasa, Făgăraş Sa1: Soroştin, Ruşi, Brădeni, Grânari, Ilimbav, Vărd Sa2: Archita, Cenade Pa: Agârbiciu, Guşteriţa, Tău
Marls	Decimeter- to meter-scale massive or laminated marl and calcareous marl. Rich in deep-water microfauna.	Pelagites, suspension fall-out on the basin plain.	Ba2: planktonic foraminifera (e.g. <i>Tenuitellina</i> spp., <i>Tenuitella</i> spp., <i>Globigerina</i> spp.) Sa1: planktonic foraminifera (e.g. <i>Tenuitellina</i> spp., <i>Tenuitella</i> spp) and rare rotallids Pa: ostracods, and molluscs	Ba2: Rupea, Făgăraş Sa1: Grânari, Făgăraş Pa: Colibi, Mihălţ, Guşteriţa, Sibiul, Bazna
<b>Evaporite</b> Gypsum	Decimeter to meter thick massive (alabaster) or thinner fibrous gypsum beds. Sometimes associated with thin calcareous marls or limestones.	Sabkha to shallow (<5 m) platform.	Absent	Ba2: Sănduleşti, Dobârca valley Sa1: Daia Română
Halite	Massive white or grey halite with very rare millimeter to centimeter thick shale intercalations; up to 300 m depositional thickness.	Deep platform (>50 m) cumulates with minor pelagics.	Absent	Ba2: Ocna Sibiu, Nireş
<b>Tuff</b> Riolite, dacite to andesite tuffs	Centimeter to meters thick beds up that form up to several tens of meter thick successions. They may be white, grey or green depending on the alteration processes.	Formed by suspension fall-out or redeposited as low-density turbidites.	Ba1: diverse planktonic foraminifera ( <i>Orbulina suturalis</i> , <i>Globigerina</i> spp., <i>Globoquadrina</i> spp., <i>Globigerinoides</i> spp.) Sa1: very rare benthic foraminifera	Ba1: Ioneşti, Perşani Sa1: Hădăreni Pa: Bazna

laminated unit, which grades upwards into an upper unit of unidirectional ripples (Fig. 4h).

#### 4.2.1. Interpretation

Wave activity is suggested by the [a(t)b(i)]-type imbrication of the pebbles from the gravel lags intercalated in the horizontally laminated sandstones. These witnessed high-energy wave activity perhaps on a low-angle foreshore where accommodation space

was limited. This could be the reason for the lack of wave ripples or dunes, but small conglomerate channels. These latter possibly represents small distributary channels incised into a sandy beach (Fig. 4a). Large-scale planar cross-stratification in sandstones may be interpreted as dunes formed by high-energy waves on the upper shoreface. Conversely, outcrops dominated by hummocky stratification and wave ripples may be indicative of the lower shoreface. This latter is in good agreement with the fauna content of the shales provided by the miliolid- and rotaliid-dominated



**Fig. 3.** Fluvial facies associations: a) Early Miocene red continental deposits at Tălmăciu (1 in Fig. 1). Sandy fluvial bar formed by lower horizontally laminated (Sh) and upper cross-bedded (St) units is overlain by an erosional channel filled with products of gravelly hyperconcentrated flows (Gmg); b) Pannonian (?) fluvial succession at Ulieș (2 in Fig. 1) with amalgamated fluvial channels. Individual channel fills contain cross-bedded conglomerates (Gp) that fining upwards into sandstones (St, Sh, Sp); c) and d) Lower Sarmatian alluvial fan dominated by hyperconcentrated flows (Gm) at Dacia (3 in Fig. 1) with massive sandstones (Sm), outsized clasts and gravel lags. The abbreviation of lithofacies for Figs. 3–6 are based on Miall (1996), as follows: G – gravel, S – sand, F – fines (shale, marls), h – horizontally laminated, r – ripple laminated, t – trough cross-bedded, p – planar cross-bedded, l – low-angle cross-bedded, m – massive, g – normal graded, i – inverse graded, l – fine laminated, and Gy – gypsum.

assemblages of foraminifera (Table 1). Overall, the common characteristics of these deposits are the presence of wave-generated sedimentary structures in sandstones (e.g. imbrication, bidirectional ripples, hummocky stratification) and shallow-marine foraminifera faunas in shales (see Table 1). Therefore, this association is interpreted as formed on the shelf and above the storm-wave base.

Massive shales with thin sandstones contain rich rotaliid and buliminid foraminifera assemblages characteristic for a low-energy shelf below the storm-wave base (Table 1). These shales are interbedded with coarse-grained, bioclastic, upwards fining sandstones and microconglomerates with massive, horizontal- and cross-laminations. The coarse-grained beds suggest reworking of littoral sediments during massive storms (Fig. 4j).

#### 4.3. Facies association III

Comprises conglomerates, pebbly sandstones, fine-grained sandstones and thin shales (e.g. Brădești, Gârbova de Jos; Fig. 1). The most common are the conglomerates and pebbly sandstones, which typically form bipartite beds represented by a lower unit of conglomerates and an upper unit of sandstones with a sharp transition (Fig. 5c).

The conglomerates are clast-supported and have erosional lower boundaries frequently marked by outsized boulders or boulder-filled erosional scours. Individual conglomerate beds form several meter thick multi-storey amalgamated units. They are covered by horizontally laminated or massive coarse-grained sandstones that are fining upwards into unidirectional ripples (Fig. 5a, b; Table 1). Imbricated [type a(p)a(i)] pebble lags are

frequent at the base of massive or horizontally laminated sandstones, forming discontinuous rows. Shale rip-ups are common scattered at different levels in the sandstones. The top of fine-grained sandstones may be bioturbated (vertical burrows).

##### 4.3.1. Interpretation

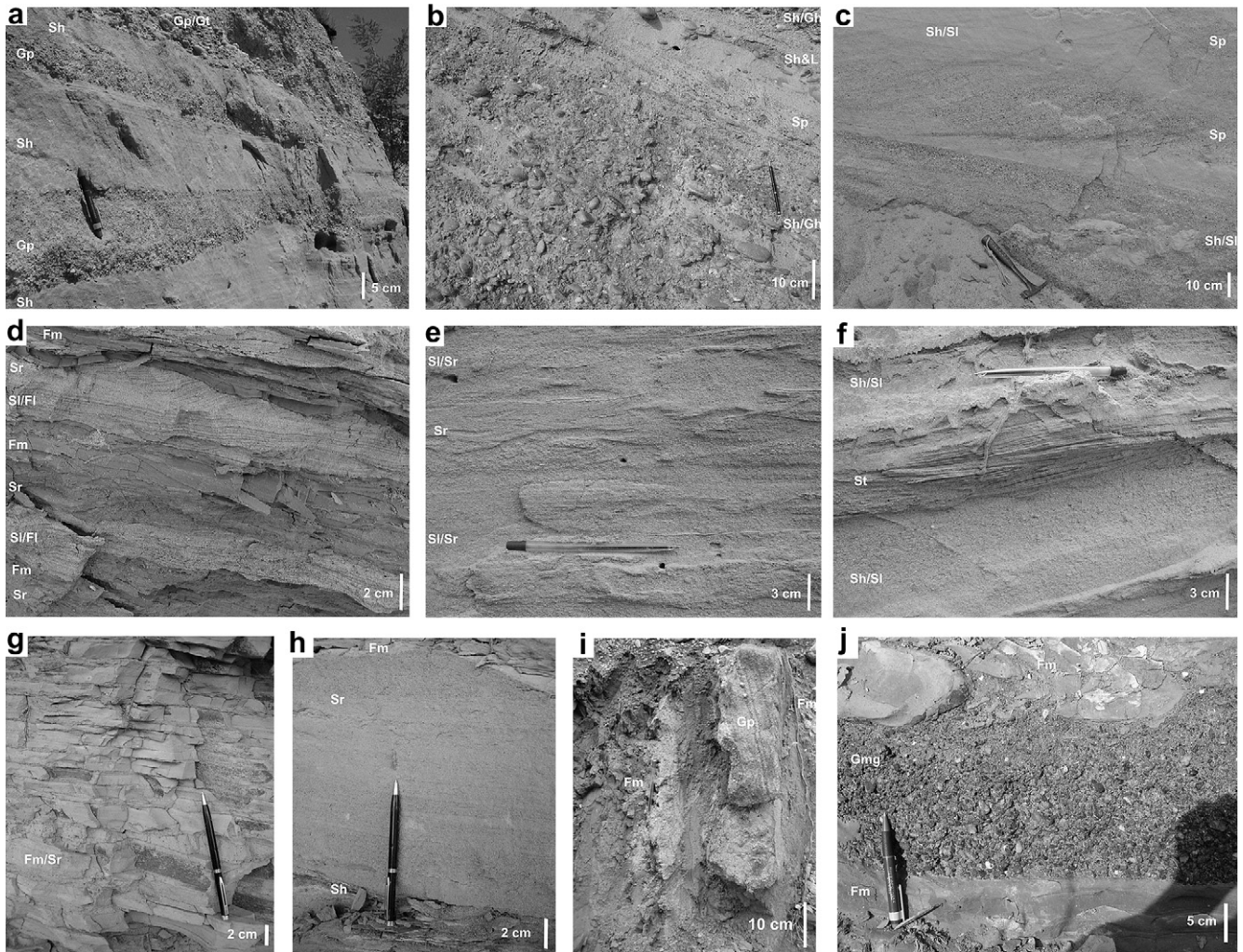
The massive conglomerates are interpreted as gravelly hyperconcentrated flows as suggested by the [a(p)a(i)]-type imbrication of the pebbles. The gravel lags and some few-clast thick boulder-sized conglomerates are considered products of debris fall. The massive sandstones that fine upwards into horizontally laminated sandstones and then into unidirectional cross-laminations are characteristic for progressively decelerating turbidites (e.g. Bouma, 1962).

The bipartite beds of conglomerates and sandstones are very similar to those shown by Surlyk (1984) and Falk and Dorsey (1998). They interpreted them as cohesionless “debris flows” (i.e. grain flows) overlain by and genetically related turbidity currents.

The beds of this facies association have up to 20° depositional dips oriented parallel to the gravity flow direction (Fig. 5d). In strike section, these successions are formed by tabular and lenticular units with highly erosional boundaries, which suggest tongue-like individual depositional units.

Overall, the architecture of this association and the depositional processes closely resemble fandelta foresets (Postma, 1990). This is suggested by the dominance of bipartite beds, high depositional dips, shallow-marine bioturbation in sandstones, shallow-marine foraminifera faunas found in the thin shales. Other indirect evidence is that commonly this association is sandwiched between various finer-grained and shallow-marine shelf deposits (see Chapter 6).





**Fig. 4.** Shallow-marine facies associations: a) Sarmatian foreshore to upper shoreface at Berghin (4 in Fig. 1), represented by lenticular planar cross-stratified conglomerates that grade upwards into horizontally stratified sandstones with scattered pebble lags; b) Sarmatian foreshore of Crăciunel (5 in Fig. 1) formed by horizontal and low-angle stratified sandstones (Sh, Sl) with imbricated gravel lags (Gh); c) Sarmatian sandy upper shoreface at Petrești (6 in Fig. 1) with well developed planar cross-stratification (Sp), horizontal (Sh) and low-angle (Sl) stratification; alternation of lower shoreface horizontal lamination, ripples and shales (d), wave ripples (e) and hummocky stratified sandstones (f) at Dobârca (7 in Fig. 1); sand-rich shelf facies association at (g) Bunești (8 in Fig. 1), (h) Mureni (9 in Fig. 1), (i) Dobârca (7 in Fig. 1) and (j) Săcădate (10 in Fig. 1) represented by sandy tempestites, often highly bioclastic in alternation of marls. See legend at Fig. 3.

#### 4.4. Facies association IV

Comprises conglomerates, pebbly sandstones, sandstones, shales and rare volcanic tuffs (e.g. Tău, Micăsasa, Archita, Rupea, Oarba de Mureș; Fig. 1). These may be grouped in various assemblages as follows: sandstones with minor conglomerates, sandstones and shales, and shales and marls with minor sandstones.

The assemblage of sandstones with minor conglomerates (Fig. 6a–e) consists of more than 80 percent of sandstones. The sandstone beds are massive and up to 5 m thick (Fig. 6b), sometimes capped by up to 20–30 cm thick horizontal and ripple laminated, finer-grained sandstones (Fig. 6c). The massive sandstones exhibit internal amalgamation surfaces that delimit individual fining-upwards units (Fig. 6d). Frequently shale rip-up clasts and highly deformed shale to siltstone beds, meters long and a few centimeters thick, occur at the basal part of the beds (Fig. 6a). The sandstones have channelised architecture often visible only on wider outcrops (e.g. Archita, Rupea, Tău). The conglomerates appear as small-scale, up to tens of meters wide channelised bodies

eroded into the massive sandstones (Fig. 6e). In general, the conglomerates represent unsorted and usually upward-fining pebble to boulder assemblages. They are mostly clast-supported, but matrix supported (mud or sand) conglomerates were found also. Clasts typically exhibit well defined [a(p)a(i)]-type imbrication in outcrops parallel to the flow direction. The conglomerates have erosional lower boundaries and sometimes grade upwards into massive sandstones. In this case, they are genetically interrelated.

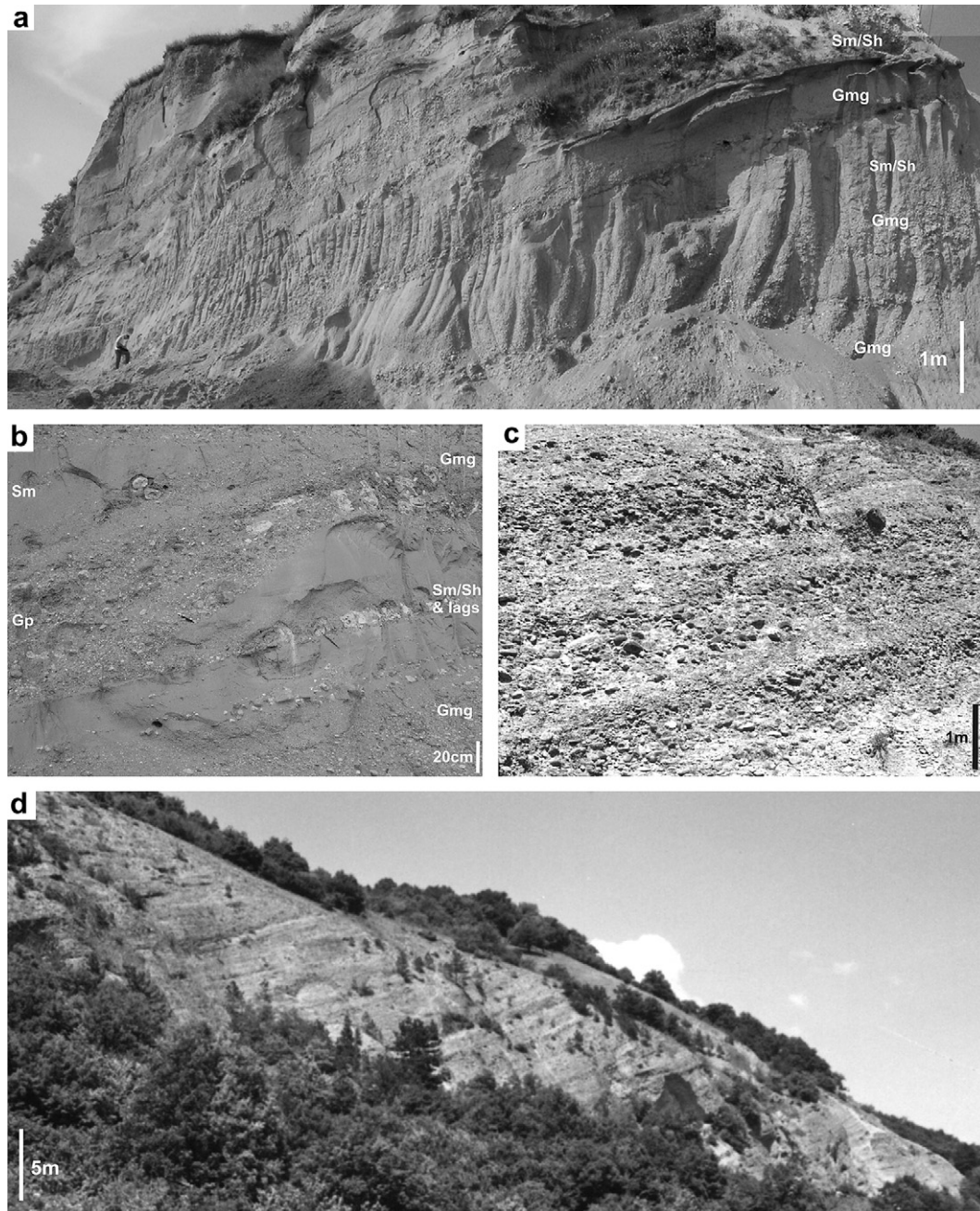
The sandstone to shale ratio in the assemblage of sandstones and shales (Fig. 6f–i) varies from 20 to 80%. Individual sandstone beds are laterally highly continuous and regularly less than 1 m thick (Fig. 6f). They have planar non-erosional boundaries (Fig. 6f) with frequent sole marks at their base. The sandstones consist of a lower massive part, rarely with small shale rip-up clasts, which grades upwards into a middle unit with horizontal lamination capped by an upper unit of unidirectional ripples and shales (Fig. 6h, i). Some of these units may be missing (Fig. 6h) or the laminations appear convoluted due to soft-sediment deformation processes (Fig. 6i). Lens-like massive sandstones, a few meters

wide, are locally present. The shales form brown to grey or whitish, a few centimeter thick layers delimited by millimeter-thick siltstone partings.

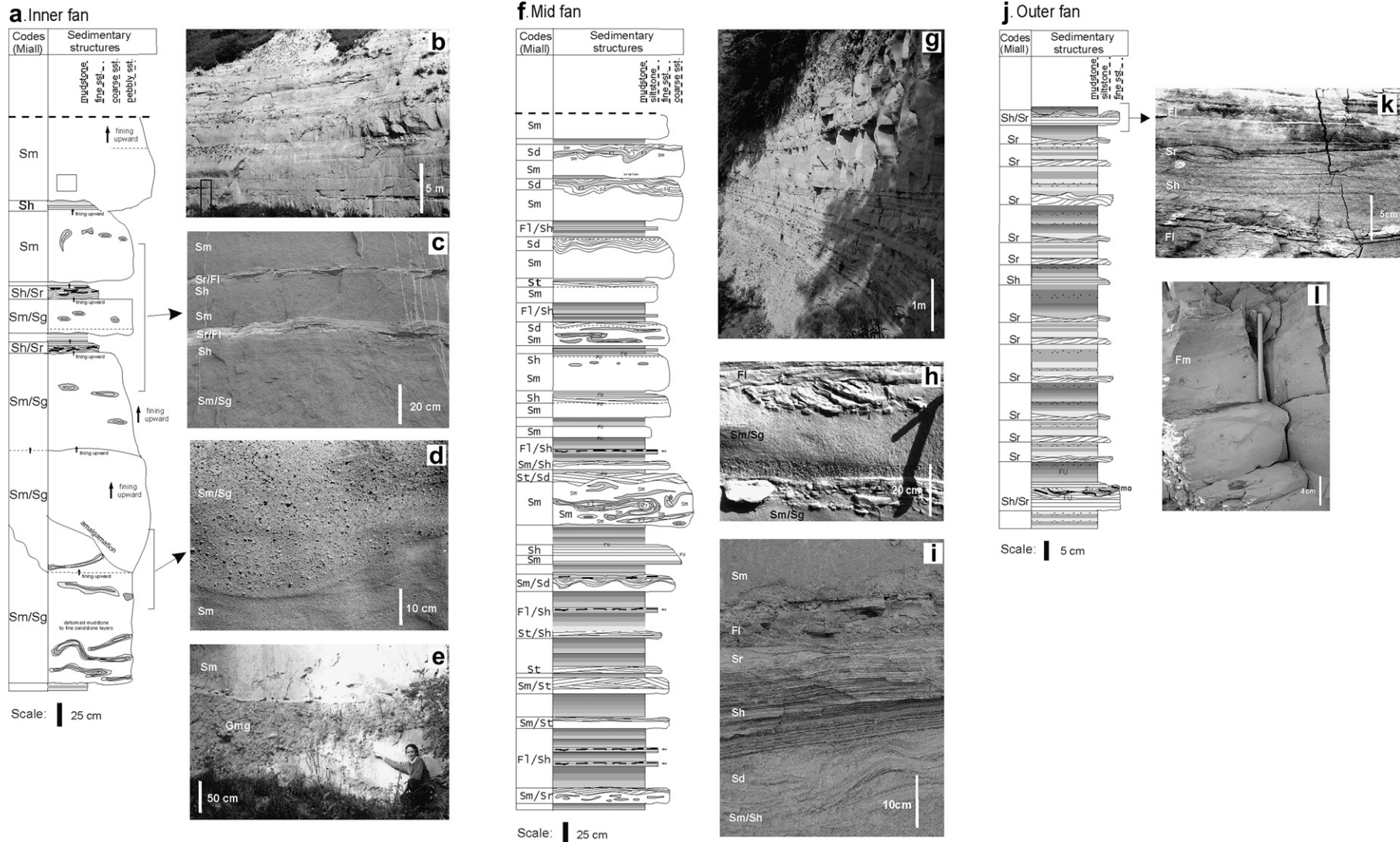
The assemblage of shales and marls with minor sandstones (Fig. 6j–l) is dominated by grayish shales and whitish marls, which form massive, up to few meters, but regularly several decimeter thick beds. Laminated shales represented by alternation of millimeter-thick shale and silt laminae appear less frequently. The sandstones may form up to 20% of this sub-facies. They are fine-grained, up to 20 cm thick and comprise a horizontal laminated unit, which grade upwards into unidirectional ripples (Fig. 6k).

#### 4.4.1. Interpretation

The [a(p)a(i)]-type imbrication in matrix supported and fining-upwards conglomerates indicate debris flows. The clast-supported conglomerates are products of hyperconcentrated flows. The massive and coarse-grained, several meters thick sandstone beds, with thin horizontal- and ripple-laminated and fine-grained sandstone caps are typical for turbidites (Lowe, 1982). Other sandstone beds, up to 1 m thick, display a succession of sedimentary structures from bottom to top of massive, horizontal-, cross-laminated units. This packages have been described as the Bouma-cycles (Bouma, 1962) formed by decelerating turbidites (Lowe, 1982).



**Fig. 5.** Fandelta facies associations: a) sand-rich fandelta foresets at Mohu (14 in Fig. 1) represented by upwards fining gravelly debrites and turbidites. Thin interbedded shales have a shallow-marine foraminifera fauna; b) sand-rich fandelta association of Călnic (11 in Fig. 1) formed by massive (Sm) and horizontally laminated (Sh) upwards fining pebbly sandstones, gravel lags and cross-stratified conglomeratic channels (Gt); c) gravel-rich Gilbert-type fandelta foresets at Brădești (12 in Fig. 1) formed by bipartite, conglomerate (Gmg) – sandstone (Sm/Sh) beds; d) birds-eye view of the gravel-rich fandelta at Brădești composed by bipartite foreset units that downlap lower Pannonian shales to the west. The fandelta foresets are part of a sand-gravel depositional system (LST8, Fig. 2a) identified only in the Eastern part of the basin. Legend is in Fig. 3.



**Fig. 6.** Deep-marine facies associations: The inner fans (a–e) consist of meters thick massive coarse-grained and pebbly sandstones (a, b) with rip-up clasts and internal amalgamation surfaces (d) that alternate with a few centimeters thick horizontally laminated and fine-grained sandstones and shales (c). Locally, a few meters wide conglomerate filled channels are incised in the sandy succession (e). The sandstones are products of turbidite currents and the conglomerates were deposited by hyperconcentrated flows. The mid fan deposits (f–i) consist of up to a meter thick medium to fine-grained sandstone beds that are massive at their base and fine upwards into horizontal laminated and then into cross-laminated units (f, i). The massive units may contain shale rip-up clasts (h). The sandstones form horizontally continuous beds with non-erosional boundaries, which alternate with meters thick laminated and massive shale packages (g). The sandstones exhibit sedimentary structures characteristic for turbidites. The outer fan deposits (j–l) are dominated by massive or laminated shales (j, l). Fine-grained horizontally or ripple-laminated sandstones occur as centimeters thick intercalations (k) interpreted as distal fringes of turbidites. The shales of deep-marine facies associations contain deep-marine fauna (Table 1). See legend at Fig. 3.

Thus, sedimentary structures of sandstones and conglomerates indicate deposition from gravity flows. These alternate with shales, which contain deep-marine foraminifera fauna (Filipescu and Silye, 2008). Therefore, this facies association is interpreted as deep-marine fans with basin plain shales. In general, the assemblage of sandstones with minor conglomerates may correspond to an inner fan represented by channelised turbidites and products of gravelly hyperconcentrated flows deposited close to the slope/basin plain break. The assemblage of sandstones and shales dominated by turbidites and minor suspension fall-out shales is inferred as mid fan deposits. Finally, the assemblage of shales and marls with minor sandstones is mainly formed by suspension fall-out of shales and rare thin-bedded turbidites. This is characteristic for the distal to outer reaches of submarine fans interbedded with basin plain fines as supported by the lack of erosional contact between the beds.

We estimate that about 70% of the preserved post-salt sediments in the basin belong to this facies association as supported by faunal evidence (Popescu et al., 1995; Filipescu, 2004a, b; Filipescu et al., 2005; Filipescu and Silye, 2008). The submarine fan systems in the basin exhibit a large variability and range from relatively small, sandy fans to larger, mixed, sandy-muddy systems (e.g. Reading and Richards, 1994; Richards and Bowman, 1998). In general, mud-rich fan systems characterize sea-level highstands, while sand-rich fans are typical for sea-level lowstands (see Chapter 6).

Volcanic tuffs are frequent in the deep-marine sedimentary successions. The tuffs are products of the Mid to Late Miocene back-arc volcanism, first active in the northern, then in the eastern part of the basin (Pécskay et al., 1995). The tuff layers are between a few millimeters to tens of meters (e.g. Ionești) thick. Most of them are developed in the western part of the basin. The only tuff layer, which covers the entire basin, is the Lower Badenian Dej Tuff (Mârza and Mészáros, 1991) and its equivalents (e.g. Merești, Perșani, Ionești). The tuffs were formed by suspension fall-out often re-deposited as turbidites (Szakács, 2003; pers. com. 2007).

#### 4.5. Facies association V

This is formed by evaporites present in two different associations: halite–gypsum (Middle Badenian, e.g. Sândulești, Praid; Fig. 1) and shale–gypsum–sandstone (Lower Sarmatian, e.g. Daia Română; Fig. 1).

The halite–gypsum association is deposited basin-wide with a minor amount of massive, nodular and fibrous gypsum. Alternations of halite and gypsum are not common and other types of intercalations (e.g. shales) are rare. The halite dominates the central parts of the basin with variable depositional thickness of up to 300 m. Few wells penetrated the salt succession. Therefore, complete evidence is rather scarce. In contrast, the outcrops on the basin western and southern margins consist entirely of gypsum. The gypsum succession is less than 20 m.

The shale–gypsum–sandstone association has been found in the southwestern part of the basin at Daia Română. It consists of centimeter-scale alternation of massive shale, fibrous or massive gypsum and horizontally laminated or massive sandstone. This association is interpreted as deposited in hypersaline lagoons.

## 5. Depositional systems and models

A depositional system (Fig. 2a) is formed by genetically linked and contemporary facies associations (Figs. 3–6). Four types of depositional systems have been recognized in the basin: 1) evaporite, 2) mud-carbonate, 3) sand-mud, and 4) sand-gravel systems (Fig. 2b–e).

### 5.1. Evaporite depositional system

The evaporite depositional system is characteristic for the Middle Badenian and is composed of shallow and deep ramps (Fig. 2a, b). The gypsum was deposited on shallow platforms (Ghargari et al., 1991) while coeval halite deposition took place in deeper areas as cumulates (Krézsek and Filipescu, 2005). Some restricted areas in the basin center functioned as morphological highs, which were likely the subject of gypsum deposition.

Evaporite occurrences have been described at large distances from the present-day basin extent (e.g. Satu Mare, Sighetul Marmației; Ciupagea et al., 1970). These suggest that the evaporite basin extended far beyond the present-day basin. In fact, the northern and eastern margins of the evaporitic basin are unknown because of the subsequent exhumation and erosion of the Carpathians (Sanders et al., 1999).

### 5.2. Mud-carbonate depositional system

The mud-carbonate depositional system is a low-energy environment, consisting of shallow-marine carbonatic or mud-rich ramps and muddy submarine fans, which is characteristic for the Lower Badenian (Fig. 2a, c).

The carbonate platforms were constructed by Leithakalk-type coralgall buildups (e.g. Gârbova de Sus; Hosu and Filipescu, 1996; Saint Martin et al., 2007). Laterally, where siliciclastic input was higher, carbonate production was restricted and siliciclastic shelves developed (e.g. Dobârca and Racoș, Fig. 4i). Some shelf sediments were reworked via slope canyons into the deep basin in mud-rich submarine fans. The fans have low sand net to gross ratios ( $N/G$ ). Their sand content on average is less than 30% (e.g. Richards and Bowman, 1998) with large amounts of reworked tuffs layers. The marls have higher proportions of planktonic foraminifera, which suggest deeper marine environments.

### 5.3. Sand-mud depositional system

This is the main depositional system in Transylvania and comprises alluvial, shallow-marine and deep-marine facies associations (Fig. 2a, d). Several deltas may have developed on rather narrow shelves. Sediments deposited on shelf were reworked by waves and further transported into the deep basin and formed submarine fans.

Alluvial deposits are known only in the Pannonian at the easternmost part of the basin (HST 8, Fig. 2a). They consist of cross-stratified fluvial gravels and sandstones, crevasse splays and overbank fines (e.g. Sâncraii).

The shallow-marine deposits crop out on wide areas near the eastern and southwestern basin margins. In well logs, they mark <10 m thick, frequently coarsening-upwards progradational cycles, consisting of facies associations ranging from distal shelf to foreshore (Figs. 4c–f, 9–HST6). Sand  $N/G$  ratios in foreshore and upper shoreface are above 90%, with low lateral continuity. Below the base of the storm wave and toward the distal shelf, the sand  $N/G$  ratio rapidly decreases to <5%.

The slope transition between the ramp and deep-water settings (Fig. 2d) can be studied at Archita (Fig. 1). The outcrop consists of a more than 30 m thick hemipelagic mudstones incised by a 15–20 m thick channel without levees, fed directly by an up dip shelf-edge delta. The channel is less than 100 m wide, but laterally is coalesced with other individual channels. The basal fill comprises conglomerate lags and debris flows, that grade upwards into turbidites, and then into mudstones. In the channel fill, the  $N/G$  ratio is around 70%. Reservoir connectivity in these sandstones is

likely to be poor because of the rapid lateral lithological changes inside and coalescent channel fills.

The main architectural elements of submarine fans are the channel/levee complexes and depositional lobes. Inner fans are dominated by channel/levees, while distal fans by depositional lobes. Particularly interesting in this setting is the channel/levee complex consisting of channels (up to 80% N/G ratio) and levees (<30% N/G ratio) facies such as the Upper Badenian at Rupea or the Pannonian at Tău (Fig. 1). These channels have a height to width ratio of 1:8 to 1:10, are lenticular in shape, erosionally based and 20–30 m in height. The lower fill is formed by amalgamated turbidites with frequent rip-up clasts. The upper channel fill is fining upwards and consists of parallel-bedded turbidites that alternate with hemipelagic muds.

The depositional lobes are made up of 2 m thick turbidites alternating with hemipelagic muds (e.g. Sztanó et al., 2005). Small, chute channels, a few meters wide can be identified. Lobes are quite frequent in outcrops and less than 20 m thick, with N/G < 50%. On well logs they display stacked bow or funnel shaped patterns. Sandstones can be laterally correlated for a few hundreds of meters at most.

#### 5.4. Sand-gravel depositional system

This gravel-rich depositional system is formed by braided alluvial systems, fandelas and sand-rich submarine fans (Fig. 2a, e). The braided alluvial systems prograded across the narrow shelves and formed gravelly shelf-edge deltas, which built fandelta foresets further on the slopes, directly feeding sandy submarine fans.

The alluvial part is poorly preserved, outcropping east of Odorheiu Secuiesc (e.g. Satu Mare, Târnovița; Fig. 1). Fandelta foresets may be examined near the basin margins. As a function of the type of the alluvial supply, the fandelta foresets may be gravel-rich or sand-rich. Typical gravel-rich foresets dipping 5–15° can be observed during the Lower Pannonian at Brădești (Fig. 5d, see also Krézsek and Filipescu, 2005), with 150 m thick package and progrades over 3 km. Along the progradation strike, individual foresets 10–20 m thick are channelised. Typical sand-rich Upper Sarmatian fandelta foresets at Mohu (Fig. 5a) are up to 100 m thick.

The submarine fans are formed by relatively elongate sandy lobes or sheets. Individual fans are up to 150–200 m thick. The well log pattern of the inner fan is sharp-based and blocky, formed by less than 60 m thick sandstone packages separated by shales (Fig. 8, LST7). The sandstone packages are progressively thinner (up to 30 m thick) and separated by thicker shales in the mid fan setting. Overall, the sand N/G ratio is more than 70% in the inner fan and decreases at less than 50% in the mid fan.

## 6. Evolution of the depositional systems

Previous interpretations of the Transylvanian Basin evolution were mostly based on geometrical correlations between major stratigraphic intervals and kinematic constraints provided by outcrop and seismic studies (Ciupagea et al., 1970; Ciulavu et al., 2000; Huisman and Bertotti, 2002; Sanders et al., 2002). We illustrate our inferences on basin evolution using seismic sections (Fig. 7) combined with generalized sedimentary trends (Fig. 8). These enabled us to construct paleogeographic maps for key time intervals (Fig. 9).

### 6.1. Lower Miocene

Near the southern basin border, ~1 km thick sequence of conglomerates at Tălmăciu (Fig. 1) has been interpreted as indicative of transgressive reworking during the Early Badenian (Ilie,

1955; Popescu et al., 1995). However, conglomerates are unusually thick for such type of deposits (package LM2 in Fig. 7b) and alluvial facies associations have been observed in the field (Fig. 3a). These sediments are younger than the underlying Lower Miocene shales (Gheorghian, 1975) and older than the overlying Lower Badenian (Ilie, 1955). If we combine these observations with their wedge-shaped geometry against the South Carpathians pre-Miocene basement and the onlaps over the pre-dating Lower Miocene shales (Fig. 7b), one can, alternatively, interpret a syn-tectonic infill, which was subsequently tilted. At a regional scale, South Carpathians underwent a Paleogene phase of rotation and exhumation linked with orogen parallel extension (e.g. Schmid et al., 1998; Fügenschuh and Schmid, 2005), which was followed in its late brittle stage by extensional/transensional collapse (Mațenco and Schmid, 1999; Răbăgia and Mațenco, 1999). Our interpretation can be therefore correlated with the later stage and the syn-tectonic wedge would be extensional. Therefore, we suggest the alluvial clastic wedge is Early Miocene in age and only its uppermost part has been reworked by the Early Badenian transgression.

### 6.2. Lower Badenian

The Lower Badenian consists of the mud-carbonate depositional system. It forms generally 1–2 reflections in seismic lines, which can be well correlated at basin scale because of its unconformable character (e.g. Dej Tuff horizon, Fig. 7a, b). Thickening patterns seem to be associated with the hanging wall of normal faults with reduced offsets. Therefore sedimentary facies, thicknesses and deformation seem to correlate, which leads us to the conclusion that at the beginning of the Badenian a limited extensional phase has affected the Transylvanian Basin. This is in agreement with similar interpretations, but less constrained in time, obtained by previous studies based on surface kinematics or seismic interpretation (Huisman et al., 1997; Ciulavu et al., 2000). We stress the limited amount of extension particularly in contrast to the coeval large-scale extensional collapse observed in the Pannonian Basin (Sachsenhofer, 1996; Tari et al., 1999) during its first stage (Fodor et al., 1999). Thus, the ~100 m Transylvanian component of synrift subsidence cannot justify in post-rift terms the 4 km sediments subsequently recorded. The scenario of a strong early subsidence and later infill of this under-filled basin is not realistic, because there are no evidences of crustal thinning in the basin, i.e. it has a thick continental crust (33–37 km) and low heat flux (26–60 mW/m<sup>2</sup>) (Visarion and Velicic, 1981; Rădulescu, 1988; Demetrescu et al., 1992).

The extension has probably led to a more complex pattern of tilted lows and highs as observed on seismic sections (e.g. Fig. 7b) than the one we depict in the paleogeographic interpretation (Figs. 2a, 9b). This is because the detailed mapping of the possibly intricate network of small extensional faults has not been attempted. In contrast, in the proximal part of the basin, the western carbonate and southern siliciclastic ramps have been clearly identified (Fig. 9b).

### 6.3. Middle Badenian

The Middle Badenian evaporite depositional system has a characteristic chaotic to transparent seismic facies. The evaporites are in general 100–500 m thick in the Southern Transylvanian Basin due to thinning over basement highs (Fig. 7). The salt forms welds, pillows, rollers and diapirs (Ciupagea et al., 1970). Pre-existing topography (the “pre-Paratethys denudational surface”, Paraschiv, 1997) highly influenced salt deposition (Fig. 7).

Salt deposition has been interpreted as a salinity crisis related to a regional fall of sea-level, which restricted the marine connections and facilitated evaporite deposition (Peryt, 2006), although this is

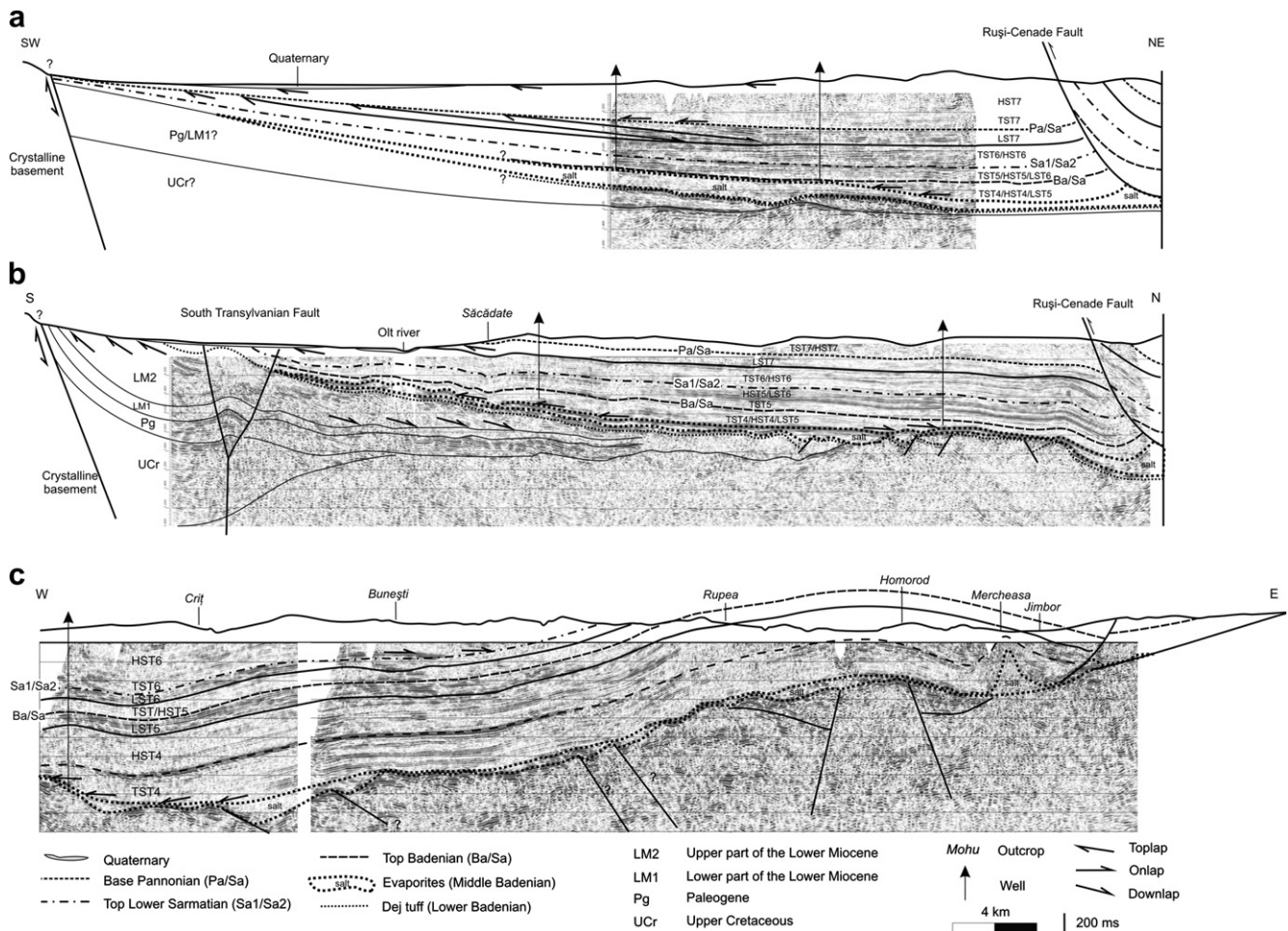
coeval with recorded tectonic related exhumation around the Carpathians arc (e.g. Tari, 1994; Csontos and Nagymarosy, 1998; Kovač et al., 1999; Krzywiec, 2001).

#### 6.4. Upper Badenian

The Upper Badenian has 1500 m in the east and decreases to less than 10 m in the west (Figs. 7, 8). This is because the Upper Badenian onlaps the top of salt and progressively pinches out to the west (TST4-HST4; Fig. 7c). Facies associations in the west and east are markedly different. Outcrops in the east consist of turbidites (e.g. Făgăraș, Comana) and pelagic mudstones (Mercheașa) with deep-marine foraminifera associations (Filipescu and Silye, 2008). They represent the deep-water part of a sand-mud depositional system (Fig. 2d). In wells (Fig. 8), the succession displays an initial fining upward (TST4) followed by a coarsening-upward trend (HST4) interpreted as transgressive and highstand systems tracts (Filipescu, 2004a; Krézsek and Filipescu, 2005). The transgressive systems tract

is unusually thick, locally more than 500 m. Its extreme thickness likely is the result of the interplay of several controlling factors, including subsidence, sediment supply and sea-level rise.

The uppermost part of the coarsening-upward trend is formed by sharp-based and blocky sandstones (LST5, Fig. 8). They correlate laterally with channelised massive sandstones (e.g. Rupea) with conglomerates (e.g. Jimbor) interpreted as turbidites and hyperconcentrated flows. Deep-marine depositional settings are indicated by the locally interbedded shales with deep-marine microfauna (e.g. Filipescu and Silye, 2008). Overall, this sand-rich package may be interpreted as an inner submarine fan, the deep-water extension of a gravel-sand depositional system (Fig. 2e). Coeval shallow-marine or alluvial deposits of such a depositional system are unknown, possibly due to subsequent erosion. The change between the sand-mud to sand-gravel depositional systems can be interpreted as a relative sea-level fall likely triggered by coeval Carpathian uplift due to nappe stacking, because no coeval eustatic event is known at either global or regional scale (LST5, Fig. 2a).



**Fig. 7.** Cross-sections in the Southern Transylvanian Basin based on seismic sections (TWT, ms) and extended using outcrop data. The location of the sections is shown in Fig. 1. a) The Upper Badenian thins to the south and onlaps the salt south of the Ruși-Cenade fault; the Lower Sarmatian is formed by shallow-marine facies associations, documented by outcrops, microfauna and well data; the Upper Sarmatian (LST7) consists of high-amplitude and downlapping reflectors, which correspond to sandy fandelas as shown by outcrop and well data (Figs. 5a, 9); the Early Pannonian is transgressive and onlaps various older units; b) The Early Miocene conglomerates (LM2) onlap Miocene shales (LM1) and wedge out to the north and thicken to the south; minor Early Badenian half-grabens may be present, but the amount of extension is very small; the salt thickens into the Early Badenian half-grabens and tapers out to the south; the Upper Badenian thins over and onlaps basement highs or salt to the north and south; the Upper Badenian outcrops at the edge of the basin consists of shelf deposits; the Lower Sarmatian is formed by shallow-marine facies associations, documented by outcrops, microfauna and well data; the lower Pannonian is made up by sandy shelf (e.g. Săcădate, Fig. 4j) and the upper Pannonian by deep-lacustrine facies associations (e.g. Gușterița). c) Tentative interpretation of the Early Badenian extension; the salt thickens and fills the Early Badenian depocenters; the Upper Badenian is deep-marine and onlaps the salt to the west; high-amplitude channels can be observed in the Upper Badenian (LST5) and Lower Sarmatian (LST6); at least two major transgressions have been recorded by the seismic onlaps in the Sarmatian; the sections evidence the westward directed gliding of the post-salt strata that has been accommodated by the growth faulting in the east mostly during the Pliocene; the eastern extensional diapirs were inverted.

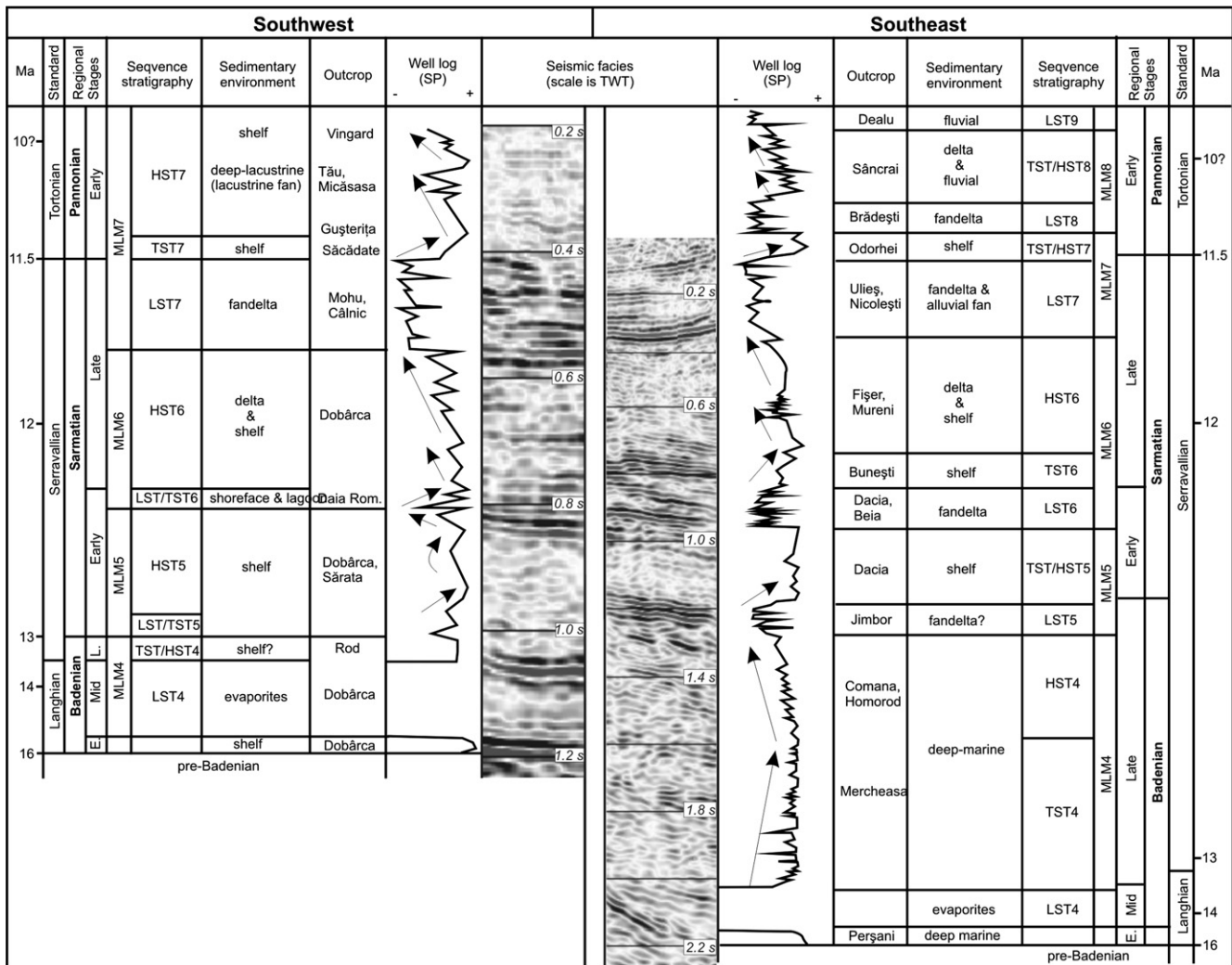


Fig. 8. Sedimentary trends of the southwestern and southeastern Transylvanian Basin based on outcrop, microfauna, well log and seismic facies correlations. The sequence stratigraphy annotation is based on Krézsek and Filipescu (2005). The location of outcrops is shown in Fig. 1. See the text for discussion.

In the west, a few meters thick Upper Badenian shales preserve deep shelf assemblages of microfauna (Filipescu, 1992). In southwest, the Upper Badenian is a thin package of shales that onlaps the evaporites or local basement highs (e.g. Fig. 7b). The shales contain deep and shallow-marine fauna (Gheorghian et al., 1967; Lubeșcu, 1981; Popescu et al., 1995). Well logs mostly consist of fining-upward trends (Fig. 8).

An E–W correlation across the basin is rather difficult. Outcrops and well logs depict the Upper Badenian deposited in deep-water conditions (Fig. 9d, e). The source of the submarine fans may have been located to the NE with depocenters being in the SE, while the basin's western part functioned as a submarine high likely dominated by shallow platform sedimentation. This is supported by the large thickness changes and differences in facies from east to west, but also by the onlaps of the eastern Upper Badenian to the west. The western submarine high of Transylvania apparently extended further the W and SW over the South Carpathians and Apuseni Mts. (Krzék and Bally, 2006).

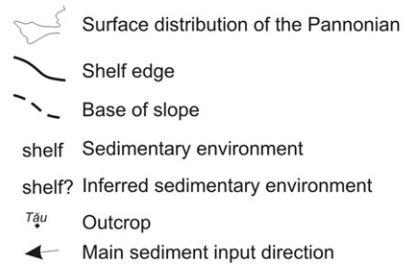
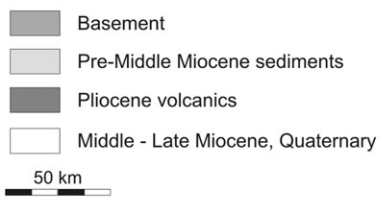
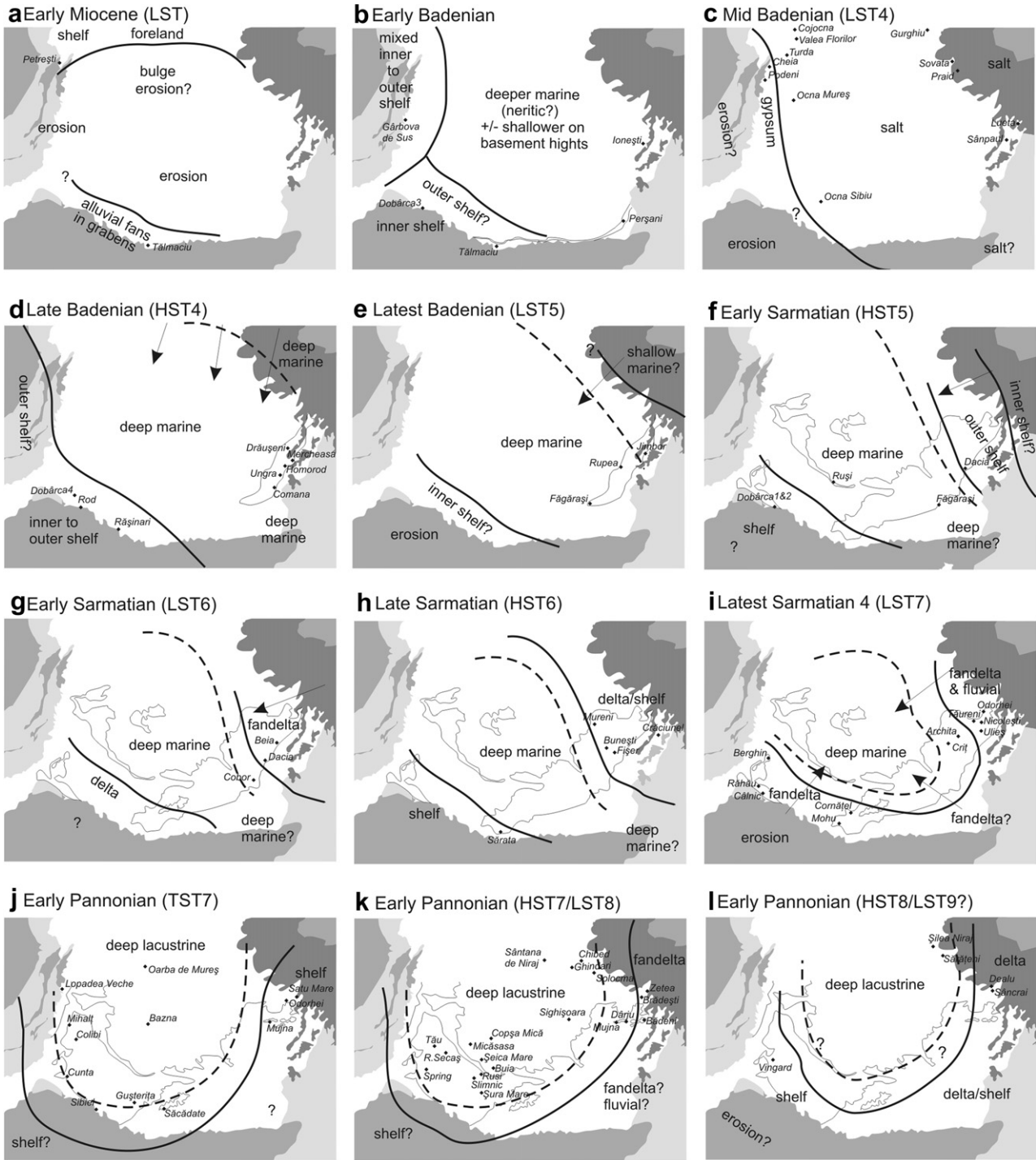
### 6.5. Lower Sarmatian

The Lower Sarmatian is a few hundred meters thick in the central part of the basin and slightly thins toward the basin margins.

It contains two different depositional systems, which are successive in time.

The lower part of the Lower Sarmatian comprises well developed shallow-marine deposits on the basin margins (e.g. shoreface sands at Dobârca, inner shelf sandstones in the Merești borehole, outer shelf marls at Dacia and Dobârca) and coeval deep-water outer to mid fan facies associations (e.g. Felmer) in the basin center. The deep-water deposits contain benthic and planktonic foraminifera of the *Anomalinoidea dividens* Biozone (e.g. Filipescu, 2004b). The well logs (Fig. 8) exhibit an early fining upward (TST5) followed by slightly coarsening-upwards trend (HST5), which record an early transgression followed later progradation (Krzék and Filipescu, 2005).

The upper part of the Lower Sarmatian consists of gravel-rich fandelta deposits (Dacia), upper shoreface sandstones capped by lagoonal gypsum (e.g. Daia Română), and inner fan sediments in the basin center (e.g. Cobor). The shale-gypsum facies association is associated with higher salinity foraminifera assemblages (*Variidentella reussi* Biozone) occasionally resedimented in deep-marine deposits. Frequently this package has a blocky well log pattern (LST6, Fig. 8) characteristic for gravel-sand depositional systems (Fig. 2e). This depositional system marks a relative sea-level fall (LST6, Fig. 2a) illustrated by the conglomerate filled incised valleys cut into the earlier highstand outer shelf deposits at Dacia. The sand





largely bypassed the former highstand shelf (HST5) using these incised valleys as transport pathways toward the basin center. In contrast, the relative sea-level fall was relatively minor in the southwest, and hence the associated erosion was rather limited (Fig. 9g). This difference indicates that the relative sea-level fall was rather tectonically controlled and related to uplift-related shortening of the East Carpathians.

### 6.6. Upper Sarmatian

The Upper Sarmatian is characterized on the basin margins by sand-rich outer shelf to shoreface facies associations (e.g. Petrescu et al., 1988) rich in shallow-marine foraminifera (*Elphidium reginum* Biozone; Popescu et al., 1995; Filipescu, 1999) and outer to mid fan facies associations developed in the basin center (Fig. 9h). The shallow-marine sediments in the southeast are up to 1 km thick and decrease to less than 200 m in the southwest (Fig. 8). The large shallow-marine systems on the basin margins and their progradational character indicate changes in sediment supply derived from the Carpathians source areas, most probably renewed tectonic exhumation.

A dramatic change in depositional systems and sedimentary facies has been recorded at the top of the Upper Sarmatian. The sediments in the east are coarse-grained, poorly sorted and gravel-rich (e.g. Sânmihai, Archita, Ulieș, Tăureni). Frequently they contain reworked shallow-marine mollusks (e.g. Nicoleşti). The thick packages of conglomerates with minor sandstones are the product of high-energy debris flows deposited in braided rivers, submarine slope canyons and inner fans. They form a gravel-sand depositional system (Fig. 9i). Erosion and locally up to 200 m of incision has been documented in the eastern part of the basin (LST7, Krézsek and Filipescu, 2005). Coeval deposits in the southwest consists of high-amplitude and continuous reflectors, which form a low-angle progradational package that toplaps the overlying Pannonian near the basin margins and flattens out into parallel reflectors in the deeper basin (Fig. 7a). In wells, this package correlates with coarse-grained sandstones that have a sharp-based blocky log pattern (Fig. 8). Microfauna from the interbedded shales belong to a shallow-marine *Porosonion aragviensis* Biozone. Outcrop analogues are made up of thick sandstones deposited by debris flows (e.g. Mohu, Câlnic, Fig. 5a).

The base level drop at the end of Sarmatian must have been produced by the tectonic uplift, due to the absence of coeval eustatic events. This is supported by the significant differences in texture between the conglomerate-rich eastern and sand-rich southwestern fandelas. The tectonic uplift was higher in the East than South Carpathians and it relates to the onset of shortening over the undeformed part of the European foreland. The uplift isolated the Central Paratethys, leading to the extinction of Sarmatian marine faunas, the basin evolving subsequently into endemic lacustrine conditions with brackish and freshwater species (Magyar et al., 1999).

### 6.7. Pannonian

The Pannonian is in general less than 500 m thick (Fig. 7). Overall, it represents a sand-mud depositional system (Fig. 2d). A local gravel-sand depositional system is observed only in the east (Fig. 2e).

The Pannonian base is observed in wells as a marly sequence with a characteristic upwards fining pattern contrasting with the gravel-rich uppermost Sarmatian (Fig. 8). This fining-upwards interval can be correlated in seismic lines with onlaps of shallow-marine deposits (Fig. 7a). These consist of inner and outer shelf deposits (e.g. Săcădate, Fig. 4j). Retrogradational coastal onlaps on top of the Sarmatian indicate a large early Pannonian transgression, which during maximum flooding established deep-lacustrine settings over most of the formerly shallow parts of the basin. This is illustrated by deep-lacustrine plain marls, which overlie the crystalline basement (e.g. Gușterita, Sibiel; Fig. 1).

In the southwest, the upper part of the Pannonian logs display stacked bow or funnel shaped patterns, which correlate to various deep-lacustrine fan facies associations (Fig. 8). These were deposited during a highstand (Fig. 2a-HST7), observed in several outstanding outcrops which illustrate the outer fan (Colibi, Fig. 6l) and mid fan (Tău) facies associations. The only shallow lacustrine succession younger than the fans was identified at Vingard (Lubenscu, 1981) with unclear relationship to the underlying deep-lacustrine deposits.

On the southeastern margin of the basin, a few hundred meters of shallow-marine and continental Pannonian has been preserved due to the cover of Pliocene volcanics, which prevented erosion. Most importantly, one major relative lake-level fall has been identified that seems to be missing from the southwest and accordingly has to be of tectonic origin (LST 8; Fig. 8). The lowstand deposits are formed by a sand-gravel depositional system represented by Gilbert-type fandelas and associated small lacustrine fans.

The uppermost part of the Pannonian is sandwiched between lowstand fandelas and the overlying volcanics (Figs. 2a, 8). The deposits are deltaic (e.g. Sâncrai) and are unconformably covered by fluvial gravels (Dealu) (Fig. 9l). The gravels are probably the only sediments preserved that witnessed the final uplift and exposure of the Transylvanian Basin. Toward the end of the Pannonian, the Transylvanian Basin was exhumed and subsequently subjected to erosion. Over 500 m of sediments were eroded (Sanders et al., 2002), in particular Pannonian deposits.

## 7. The Miocene petroleum system

### 7.1. Exploration history

Surface gas emanations and “burning” wells of Transylvania have been mentioned in various manuscripts starting with the 17th

**Fig. 9.** Paleogeographic reconstructions for selected Miocene systems tracts of the Southern Transylvanian Basin. a) During the Early Miocene foreland basin develops in the north. Conversely, the southern part of the basin is influenced by the transtensional collapse of the Southern Carpathians (grabens with alluvial fans). The central areas are eroded; b) Early Badenian transgression establishes marine settings all over the basin. Weak extension recorded. Carbonate ramps dominate the western margin and muddy shelves and deep-marine settings other areas (mud-carbonate depositional system, Fig. 2c); c) Mid Badenian salinity crisis triggered salt (halite) and gypsum deposition (Fig. 2b); d) increased subsidence during the Late Badenian establishes extensive deep-marine environments dominated by sand-mud depositional systems (Fig. 2d); sediments are shed into the basin from the uplifting north Eastern Carpathians; e) relative sea-level fall by the end of the Late Badenian induce lowstand conditions witnessed by extensive reworking of shallow-marine strata and sand-rich lowstand fans; f–i) Shallow-marine environments are firmly established with the onset of the Sarmatian. Highstands of relative sea-level (f, h) are characterized by the sand-mud depositional system (Fig. 2). This contrast with periods of relative sea-level lowstand (g, i) distinguished by the sand-gravel depositional system and associated coarse-grained fandelas (Fig. 2e). The Latest Sarmatian lowstand (i) marks the final continental collision and major uplift of the Eastern Carpathians. This later event induces restriction of open marine seaways that progressively transforms the marine basin into a large freshwater lake (Magyar et al., 1999); j–l) relative lake-level rise at the base Pannonian transgresses the Latest Sarmatian alluvial and coast plain settings and develops narrow lacustrine shelves with deltas fringing the basin margins. Sandy lacustrine fans in the basin center are line-sourced by several shelf-edge deltas. Tectonic uplifts in the hinterland are marked by local relative lake-level falls that are associated with progradation of gravelly Gilbert-type fandelas (k); During the Late Pannonian the basin is uplifted and eroded. No Upper Pannonian sediments are preserved, but some fluvial gravel at Dealu (38 in Fig. 1) where the Pliocene volcanic cover prevented erosion.

century (see Wanek, 1999). Systematic exploration efforts however started only in 1909. By then an exploration well prospecting for potassium salts accidentally erupted when it encountered a gas reservoir (Paraschiv, 1979).

Early exploration consisted of fieldwork campaigns aimed to map 4-way dip closures (“gas domes”; Böckh, 1911; Mrazec and Jekelius, 1927; Vancea, 1960; Ciupagea et al., 1970). Later, by using seismic methods in exploration other play types have been identified resulting in the successful drilling of structures. All discovered gas fields had several pay zones exclusively located in the post-salt sedimentary succession (Fig. 2a).

The pre-salt deposits have been explored with more than 40 wells drilled by Romgaz and later, in the 1990s, by Shell. Exploration targeted different plays in Jurassic carbonates, Upper Cretaceous turbidites and Eocene shallow-marine reservoirs, but all were tested dry (Paraschiv, 1979). The only exception is the 6042 Deleni well, which had oil shows in highly cemented Jurassic dolomites (Kovács et al., 2007).

Geophysical methods (e.g. gravity, magnetic, and 2D seismic surveys) were applied in exploration since the 1950s. Early data quality was rather poor, but good enough to identify the main structural features and visualize major unconformities. Good quality 2D seismic became available in the 1990s and since then exploration has involved the acquisition of dense grids of 2D seismic data over post-salt structural plays.

Seismic interpretation is typically done on seismic sections processed in time, because regional velocity models are not well established. No 3D seismic surveys have been used in exploration so far. In fact, the first 3D seismic survey has been shot a few years ago for field appraisal purposes.

Exploration was very successful until the 1980s, but subsequently recorded a continuous decline (Paraschiv, 1997). More than 130 gas fields were brought to production. Gas field sizes range between 6 BCF to 1 TCF, with the great majority of fields rather small, less than 100 BCF (Pawlewicz, 2005). The Romanian state-owned gas company (S.N.G.N. Romgaz S.A.) operates most of the gas fields and the exploration licenses in the basin. Gas production started in 1913, peaked at 1 TCF in 1976 then progressively dropped to 212 BCF recorded in 2007. Overall, more than 40 TCF gas has been commercially produced (de Broucker et al., 1998). Production from most gas fields is declining with often only 30% of the initial reserves left (Nemeşan, 2007).

### 7.2. The origin of the Transylvanian dry gas: biogenic vs. thermogenic

The gas in Transylvania is dry and comprises more than 99% of methane and minor amounts of ethane, propane, CO<sub>2</sub> and N<sub>2</sub> (Paraschiv, 1979). This is characteristic for more than 90% of the gas fields. Minor amounts of condensate (<1% butane) mixed with methane were found in the Upper Badenian exclusively above the Upper Cretaceous depocenters (e.g. Krézsek and Bally, 2006). Unusual high CO<sub>2</sub> (up to 99%) and N<sub>2</sub> (up to 40%) concentrations typify gas fields situated in the close proximity of Pliocene volcanics at the southeastern borders of the basin (Paraschiv, 1980). This is a possible effect of the volcanic activity, which provided additional heat and gases to the hydrocarbon system.

The origin of the dry gas is uncertain because modern isotopic measurements and source rock to hydrocarbon correlations have never been published. Constraints on the possible source(s) of the dry gas may be provided by 1D thermal models (Crânganu and Deming, 1996; Morariu, 1998) and some other considerations as follows.

Crânganu and Deming (1996) presented 1D thermal models and sensitivity tests, which capture various burial and heat flux histories of the basin (Fig. 10). Similar results are presented by Morariu (1998).

The basin has a low heat flow, in average 45 mW/m<sup>2</sup>, and ranges from 26 mW/m<sup>2</sup> in the basin center to 50–60 mW/m<sup>2</sup> to its margins (Demetrescu et al., 1992; Andreescu et al., 2002). The low heat flow fits well with the presence of a 100 km thick and unextended lithosphere and the Moho depth at 34 km (Tiliţă et al., 2006).

All thermal models predict thermally immature post-salt deposits, which fits well with the Rock-Eval analysis of post-salt shales (e.g. Popescu, 1995). Thermal maturity was reached in the Late Miocene by pre-salt deposits deeper than 4–5 km (Fig. 10). Thus, the deepest Cretaceous and Upper Jurassic may have reached the oil window and even the dry gas generation phase. This is supported by the oil shows in Upper Jurassic carbonates encountered by only one well in the entire basin (6042 Deleni; Bucur et al., 2004). Unfortunately, no direct evidence exists for thermal maturity of the Upper Cretaceous or older deposits buried deep in the basin. The Upper Cretaceous that crops out at the western border of the basin has Vitrinite Reflectance values up to 1.2 (Schuller, 2004). Pre-salt source rock candidates are the Upper Cretaceous shales and the Upper Jurassic radiolarites (Ciupagea et al., 1970; Bucur et al., 2004). These have a restricted development in the basin (Krzék and Bally, 2006) and their source properties have not been studied.

An alternative scenario for generation is the dry gas originates in the post-salt shales by biogenic decomposition process (e.g. Popescu, 1995). The composition of the Transylvanian dry gas with more than 99% of methane is typical for bacterial gases (e.g. Clayton, 1992). Bacterial decomposition of organic material occur in basins with sedimentation rates of 200–1000 m/My at depths less than 3 km and geothermal gradients of 20–40 °C (Clayton, 1992). Source rocks for bacterial gas may have low, less than 1% TOC content (e.g. Kotarba et al., 1998; Schultz et al., 2009). All these fit well the characteristics of the post-salt Mid to Late Miocene Transylvanian Basin.

In conclusion, the dry gas may be thermogenic sourced from the deep Mesozoic or biogenic sourced from post-salt shales or both. We favor the latter, because the lack of butane and condensates from the dry gas, but in fields situated above the Upper Cretaceous depocenters. However, more data is needed to better define the origins of the dry gas.

## 8. Discussion: inferences on the petroleum system elements

The key elements of the Transylvanian petroleum system are illustrated in Fig. 11. The Mesozoic petroleum system (e.g. Popescu, 1995) is beyond the scope of the present paper, but we include here to show the marked difference between the two petroleum systems.

In the following, we are going to discuss in detail the Miocene petroleum system including the distribution and quality of the source rock candidates, reservoir and seal rocks, the trap types and their timing relative to the migration phases. These enable us to recognize different play trends in the basin and to address some of the risks associated with them.

### 8.1. Source

Post-salt source rock candidates are the Upper Badenian and the Sarmatian shales (Figs. 2a, 11). These have a 0.5–1.2% TOC content, formed by Type II and Type III kerogen, S<sub>2</sub> 0.14–0.48 mg HC/g, T<sub>max</sub> 423–436 °C (Popescu, 1995; Morariu, 1998). The shales are gas-prone, have a low source potential and are thermally immature (Popescu, 1995). The best source rocks (around 1% TOC) are the lower Upper Badenian and lower Sarmatian deep-marine shales, part of sand-mud depositional systems. Their source potential is a result of sedimentary condensation developed during regional transgressions (TST4&5, Fig. 2a). Among these, high organic content

is associated with the Upper Badenian laminated shales, associated with restricted deep-water circulation and anoxic environments. Source rocks with low organic content are the deep-marine shales formed during highstand systems tracts (HST4–6, Fig. 2a). They contain large amounts of Type III land-derived organic material shed by deltaic systems into the deep basin.

The presence and quality of the Upper Badenian source rocks as supported by the paleogeographic reconstructions (Fig. 9d, e) indicate that the laminated shales are thick in the central and eastern parts of the basin, are thinning to the west and most likely missing from the SW. The advance of Sarmatian sandy deltas especially from the east (Fig. 9g–i) drastically reduced the presence and quality of source rocks in the basin. The central part of the basin remained the best area for source rock deposition. Additionally, the shales deposited on the outer shelf (Fig. 9f, h) might have a rather low source potential.

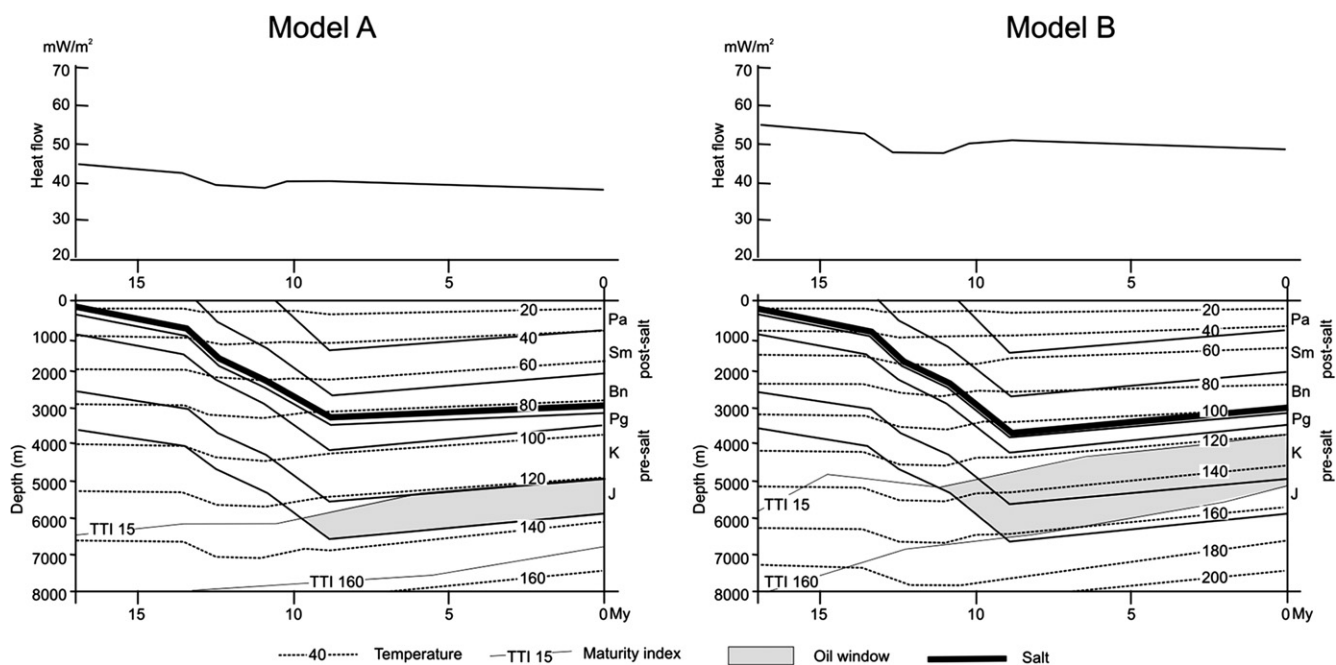
## 8.2. Reservoir

The best reservoirs are found in the central parts of the basin (Fig. 9e, g, i) represented by multi-storey sheet-like sandstone beds correlable for several hundreds of meters as shown by producing fields (e.g. Paraschiv, 1979). Individual sandstone beds up to 10 m thick have mean porosity values of 15–30% and permeabilities of 20–1000 mD (Ciupagea et al., 1970, Paraschiv, 1979). Upper Badenian reservoir quality significantly decreases below depths of 2.5–3 km, where permeabilities are less than 100 mD (Paraschiv, 1979). According to this study, the reservoirs from the central parts of the basin are deep-marine turbidites, part of mid fan, and in lesser degree inner fans (Facies Association IV, see Section 4.4). Analogues for mid-fan reservoirs are the outcrops of Comana (Late Badenian; Fig. 6g), Cenade (Sarmatian) and Copșa Mică (Pannonian). There,

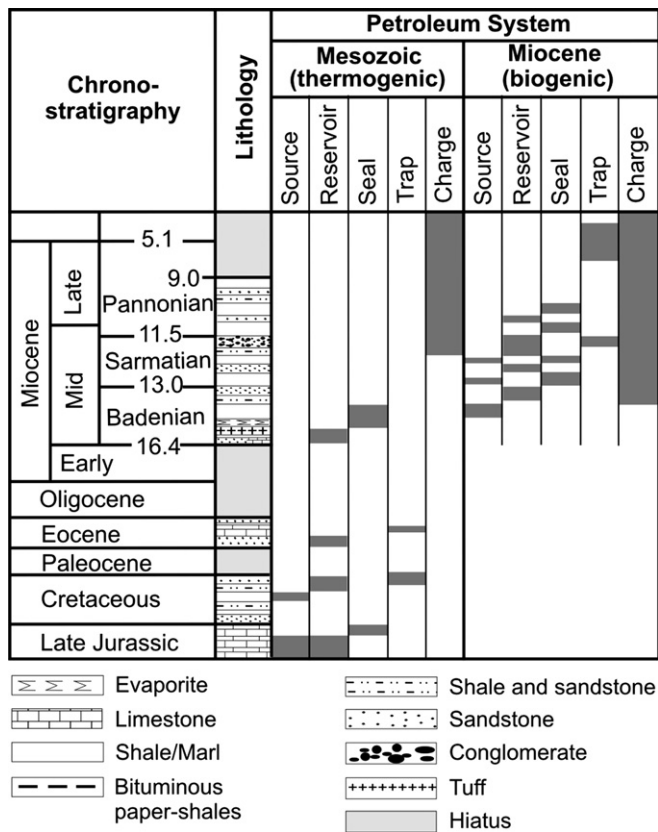
a few meters thick turbidites alternate with shales. The sandstone and shale beds are laterally continuous, correlable for hundreds of meters (e.g. Copșa Mică) without evidence of major changes in thickness and facies. In this context, the thick shale packages may act as intraformational seals and limit the vertical connectivity of sandstones. This is commonly seen in many producing reservoir packages built by several individual pay zones (Paraschiv, 1979).

Vertical connectivity is good in the inner fans because of the lack of thick and laterally continuous shales, but in the uppermost channel fill as shown at Rupea (Upper Badenian) and Tău (Pannonian). The inner fan sandstones are good reservoirs, but relatively limited in size compared with the sheet-like sandstones of the mid fans. For example, the inner fan channel at Tău is about 300 m wide and 30 m deep filled with massive sandstones. This channel has no conglomerates, which are common at the eastern part of the basin. The presence of lenticular conglomerates interbedded with massive sandstones increases reservoir heterogeneity and decreases reservoir quality, because the conglomerates are characterized by low permeability (1 Darcy) and porosity (<10%) as evidenced by several wells drilled in the central-eastern part of the basin (Ciupagea et al., 1970). Outcrop analogues for conglomerate-rich inner fan channels are at Jimbor (Upper Badenian) and Ghindari (Pannonian). These channels typically characterize the relative sea-level lowstands, in particular those of uppermost Badenian (LST5), lower Sarmatian (LST6) and uppermost Sarmatian (LST7) (Fig. 2a; Krézsek and Filipescu, 2005).

Slope channel systems developed in the northern and eastern parts of the basin are rich in conglomerates as indicated by wells and outcrops (e.g. Archita) with porosities < 10% and permeabilities < 1 Darcy (Ciupagea et al., 1970; Paraschiv, 1979). The channel system at Archita has 80% sands and 20% conglomerates. Another channel, which crops out at Nicoleşti (Fig. 1), is conglomerate-rich with 60%



**Fig. 10.** Thermal models of Crânganu and Deming (1996; their Fig. 5e, f) modified to correct for the sedimentary thicknesses characteristic for the basin center (thinner Paleogene, thicker Badenian) and to include the up to 500 m of erosion in central Transylvania that occurred due to the post-9 My uplift of the basin (Sanders et al., 1999). The late heat flux of the basin has been lowered with 2.5 mW/m<sup>2</sup> to account for the uplift of the basin according to the estimates of Andreescu et al. (2002). The most likely model is Model A with average heat flow (35–40 mW/m<sup>2</sup>). Model B has high heat flow (40–50 mW/m<sup>2</sup>). In both models, the Paleogene and Miocene are thermally immature. Oil generation in the Jurassic starts at 11 My in Model B and at 7 My in Model A. This is due to the heat flow difference between the models. Note, that the Upper Cretaceous is mature in Model B and immature in Model A. The thermal models fit well with the shows of thermogenic gas and oil found in the Jurassic and Upper Cretaceous, but fail to explain the vast amount of dry gas exclusively found in the post-salt succession (see discussion in Chapter 7.2). We support earlier interpretations of bacterial origin of the Transylvanian dry gas. Abbreviations: Pa – Pannonian, Sa – Sarmatian, Bn – Badenian, Pg – Paleogene, K – Cretaceous, J – Jurassic.



**Fig. 11.** Summary of the Mesozoic and the Miocene Petroleum Systems. The Mesozoic Petroleum System follows Paraschiv (1979), Popescu (1995) and Ciulavu et al. (2000). Source rock candidates are limestones (Late Jurassic) and deep-marine shales (Late Cretaceous). The best reservoirs are shallow-marine sandstones (Eocene) and turbidites (Late Cretaceous). These may be trapped in Late Cretaceous structural closures or Eocene stratigraphic traps frequently sealed by shales (Early Badenian) or salt (Mid Badenian). The thermogenic charge timing is Sarmatian to present-day. No commercial accumulations of hydrocarbons, but shows have been proved in the Mesozoic Petroleum System. The Miocene Petroleum System comprises large amounts of biogenic gas sourced from Middle Miocene shales (e.g. Late Badenian, Early Sarmatian). The gas frequently is trapped in deep-marine turbidites (Late Badenian, Sarmatian) located in large 4-way dip closures. Most structural traps are related to salt-cored folds that show a weak growth phase during the Sarmatian, which is followed by a strong post-Pannonian inversion. The charge has started in the Badenian and is still active today. For discussions please refer to text.

conglomerates and 40% sandstones. Individual channels are less than hundred of meters wide and 10–20 m high and coalesce laterally on the slope to form up to a kilometer wide amalgamated channel systems (e.g. Archita). The high percentage of conglomerates significantly reduces reservoir quality (Paraschiv, 1979).

Sarmatian shallow-marine sandstones in the eastern and southern parts of the basin produced with moderate flow rates, but only for very limited period. This is because sandstone connectivity is low, due to the significant lateral and vertical facies changes as observed at outcrops (e.g. Dobârca, Satu Mare, Buneşti). These facies changes are difficult to predict using 2D seismic and sparse well data.

In general, the western and southern parts of the basin are poor in sands (Figs. 2a, 9f, h; Ciupagea et al., 1970). Thin-bedded reservoirs encountered by some wells produced with low flow rates (e.g. Sâncel, Tăuni, Paraschiv, 1979). This is because the reservoirs consist of centimeter-scale intercalations of fine sands in thick mudstones interpreted as distal outer fan turbidites interbedded with pelagic shales.

Pannonian deep-lacustrine sandstones have good reservoir properties (Paraschiv, 1979), but are rarely saturated with gas (e.g.

Sădinca, Alămor, Sângeorgiu de Pădure), because they lie at shallow depths (less than 300 m) and many times lack top seals. The Pannonian is not considered a play in the basin, but in the East where it is overlain by volcanics.

### 8.3. Seal

The seals are represented by shales (Fig. 11). The sealing properties of the shales (e.g. brittleness index, capillary entry pressure, etc.) are not known, because to our knowledge no detailed measurements have been ever performed. Pay zone thicknesses indicate that the seals can hold up to 60 m gas columns. Typically, the encountered gas columns are less than the maximum closure height of the traps. This leads us to suppose that the 60 m column height is close to their sealing capacity. The brittleness of the seals was enhanced by the late uplift of the basin and most likely the seals are brittle in the basin. The presence of multi-storey stacked pay zones at normal hydrostatic pressure in many gas fields may be an indirect evidence of rather poor and brittle seals, which leak off vertically when the gas columns exceed 60 m.

The best seals are the lower Sarmatian and lower Pannonian shales (Fig. 2a). This is because during the early Sarmatian and early Pannonian several tens of meter thick shale packages were deposited mainly in the inner part of the basin due to the transgressive and early highstand conditions (Fig. 2a). These shales optimally seal the lowstand fans developed by the Late Badenian (LST5) and Late Sarmatian (LST7) (Krezsek and Filipescu, 2005) and typically form the best plays in the basin (Section 8.6; Ciupagea et al., 1970). Other good seals may develop during highstand systems tracts (e.g. HST5, HST6), but these are rather local, intraformational seals.

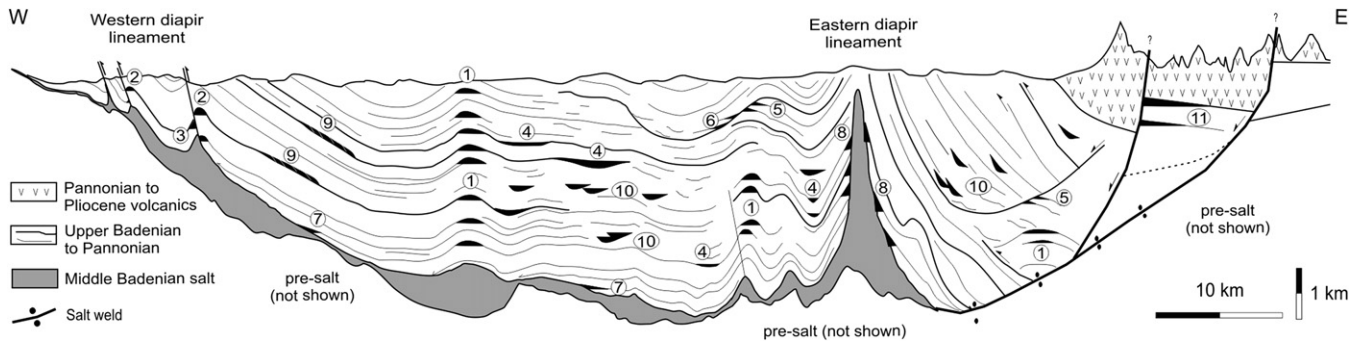
The western and central parts of the basin have a low seal risk given by the presence of tens of meter thick shale packages. Moderate seal risk characterizes the flat-lying, shallow-marine Sarmatian rich in sand in the southeast and southwest, due to interbedded shales which are laterally limited and change frequently the depositional facies. High seal risk occurs near the basin margins, where their tectonic uplift has tilted and subsequently eroded sediments.

### 8.4. Traps

#### 8.4.1. Structural traps

The most common structural traps are the 4-way dip closure associated with salt-cored folds (1 in Fig. 12). This typifies the vast majority of the gas fields in the basin center (e.g. Nadeş, Deleni, Sărmaşel, Cetatea de Baltă, Târgu-Mureş, Filitelnic, Bazna, etc.). The anticline flanks dip between 1 and 10° (Ciupagea et al., 1970). Frequently, the large 4-way dip closures are heavily compartmented due to the complex radial network of extensional (key-stone) faults formed by the releasing bending stresses during the folding (e.g. Deleni). Secondly important is the 4-way or faulted 3-way dip closure linked to the hanging wall of reverse faults (2 in Fig. 12). This is characteristic for the western part of the basin (named “Western Diapir Lineament”), locally in the east (e.g. Feliceni) and along the Ruşi-Cenade fault in the south (e.g. Ruşi) (Fig. 1). The latter is a shallow reverse salt décollement thrusting Sarmatian on top of Pannonian strata (Fig. 7a, b). A rarely drilled and poorly known trap is the footwall sub-thrust closure beneath the reverse faults (3 in Fig. 12). This has been tested at the Ruşi-Cenade fault, but no significant gas accumulations have been found yet. No data is available about the sealing capacities of the faults or the degree of fracturing of the fault zone.

The closures typically contain up to 15 vertically stacked individual pay zones. Each pay zone may be up to 60 m, but frequently less than 25 m thick and has different gas-water contacts (e.g. Paraschiv, 1979). Water saturation in reservoirs varies between 30 and 60%. The formation water contains 30–95 g/l CaCl<sub>2</sub>, MgCl<sub>2</sub>



**Fig. 12.** Trap types (in black) of the Transylvanian Basin schematically illustrated on east to west regional cross-section based on Krézsek and Bally (2006). The cross-section location is shown in Fig. 1. The most common traps are 4-way dip closures in salt-cored folds (1) or hanging wall of reverse faults (2). Other plays the sub-thrust closure associated with reverse faults (3) or faulted 3-way dip closure. Stratigraphic traps are confined submarine fans (4), toplaps (5) and onlaps (6) on unconformities, onlaps on top salt (7) and diapir flanks (8), forced regressive deltaic wedges (9), and lowstand canyons and incised valleys (10). Stratigraphic and/or structural traps beneath the volcanics characterize the eastern part of the basin (11).

and  $\text{NaHCO}_3$ . Most pay zones are at normal pressure, close to the hydrostatic (0.1–0.115 atm/m), but slight overpressure (0.12–0.285 atm/m) has been recorded as well (Ciupagea et al., 1970). Individual pay zones recorded early production of up to 170,000  $\text{m}^3$ /day, but often less than 50,000  $\text{m}^3$ /day (Paraschiv, 1979).

The structural traps are part of north to south trending salt-cored compressional folds, which exhibit quite regular wavelength in the basin (10–15 km, Fig. 1). The folds detach on the salt, i.e. the folding does not affect the pre-salt deposits, and cease southwards at the Ruși-Cenade fault (Fig. 1; Krézsek and Bally, 2006). Kinematic analysis of the syn-growth strata of the folds or the hanging wall of the reverse faults suggest weak deformation during the Late Badenian to Sarmatian and strong post-Early Pannonian shortening due to a main NE to SE oriented principal stress (Ciulavu et al., 2000). The latter has been interpreted as an effect of the westward gliding of the post-salt succession off the uplifting Eastern Carpathians (Krézsek and Bally, 2006) or reflect the intra-plate deformation observed in the whole intra-Carpathian area (Bada et al., 2001; Bertotti et al., 2003). Thus, most structural traps formed late, post-9 My in the basin. By that time, the exhumation of the basin and related erosion started. Consequently, the vast majority of structural traps require a late charge.

#### 8.4.2. Stratigraphic traps

The stratigraphic traps have never been defined in Transylvania (Fig. 12). The potential for stratigraphic traps is hinted by the structural traps common in the eastern part of the basin (e.g. Lunca, Voivodeni, Corunca-Sânișor, Cristur, Târcești, Lupeni), which reveal complex internal architecture with frequent pinch outs of beds and rapid changes of facies (Paraschiv, 1979).

Many stratigraphic traps are associated with syn-sedimentary salt tectonic processes (e.g. Krézsek and Bally, 2006). Early growth of salt-cored folds created highs on the basin floor, which constrained deposition of submarine fans in the intervening lows (4 in Fig. 12). Therefore, some fans pinch out on local highs (e.g. Deleni-Hărănglab, Zau de Câmpie – Șăulia; Ciupagea et al., 1970). These may be attractive exploration targets in particular where sand-rich lowstand fans are sealed by transgressive and early highstand shale packages as frequently occurs in the uppermost Badenian and the Sarmatian.

Major unconformities are associated with salt tectonics in particular in the Sarmatian at the eastern part of the basin. Tilted and highly eroded beds toplap the unconformity, while younger beds onlap it (5 and 6 in Fig. 12). Both strata terminations are promising exploration targets as shown by the recent discoveries of gas in toplap trap at Corunca-Sânișor. Onlaps formed by the Upper

Badenian on the salt may represent interesting exploration objectives if sands are encountered (7 in Fig. 12).

Diapirs more than 3 km high dominate the eastern part of the basin (named “Eastern Diapir Lineament”, Figs. 1, 12). The diapirs witnessed passive growth (e.g. Vendeville and Jackson, 1992) during an early extensional phase (Late Badenian to Sarmatian) and subsequently inverted in the post-Pannonian (Krézsek and Bally, 2006). The Upper Badenian and Sarmatian onlaps the diapir flanks and possibly forms traps that are typical for many salt basins around the world (8 in Fig. 12).

Relative sea-level falls (e.g. LST6–8 in Fig. 2a) triggered development of forced regressive deltas (9 in Fig. 12) and incision of alluvial valleys on the shelf, canyons on the slope (10 in Fig. 12). These may function as stratigraphic traps if a proper seal is present.

The sub-volcanic play (11 in Fig. 12) is untested in the basin. Only a few sub-volcanic wells have been drilled in poorly understood structural settings. We speculate that the volcanics may function as seal for the Pannonian deposits below and form stratigraphic traps (e.g. Rosebank field in West Shetlands; Helland-Hansen et al., 2007). No studies have been done to assess the sealing capacity of the volcanic shales. Combined structural (e.g. rollover anticlines) and stratigraphic (e.g. channels) traps may be present at deeper levels related to the gravitational spreading of the post-salt overburden (Krézsek and Bally, 2006).

#### 8.5. Migration

Structural traps have been affected by the late shortening and inversion of the basin that started at around 9 My (Fig. 11). This may be shown by flattening a seismic section on a Mid Miocene horizon (e.g. Fig. 12), which indicates that most of the pre-9 My traps formed only very gentle low-amplitude folds (Krézsek and Bally, 2006). Therefore most of the migration in the traps must have been occurred very late, post-9 My.

In addition, uplift and erosion of the basin margins occurred along with the rising Carpathians (Ciulavu et al., 2000). This generated a basement slope of up to  $5^\circ$  to the east and to the north (Krézsek and Bally, 2006). This tilt likely caused migration of the gas from the central parts of the basin toward the margins, in particular to the north and to the southeast where the amount of uplift is the highest (Ciupagea et al., 1970; Sanders et al., 1999). Similar uplift is recorded on the eastern margins of the basin. There the volcanic cover might have prevented the escape of the gas. The process of gas migration and escape is still ongoing as indicated by several surface gas emanations (Baciu et al., 2007).

### 8.6. Play trends and exploration risks

To assess the exploration risk associated with different areas of the basin we combine the earlier described source, reservoir, seal and trap evaluations (Fig. 1). The central part of the basin has the lowest hydrocarbon system risk, as shown by the large number of existing fields (Area 1, Fig. 1). The reservoirs are deep-marine sandy turbidites (mid fan facies associations) sourced by transgressive and highstand deep-marine shales (Fig. 2a; Krézsek and Filipescu, 2005). The lowstand fans are well sealed by thick transgressive and highstand shale packages. The four-way closures have several vertically stacked pay zones (Fig. 12). A dense 2D seismic grid is available, and large non-drilled closures are unlikely to still exist. Exploration potential consists of detecting smaller closures on 3D surveys possibly flanking already known fields. The main potential lies in stratigraphic traps represented by confined basin floor fans that pinch out on salt-cored folds.

The hydrocarbon system of Areas 2 and 3 is not as good as in the central parts of the basin (Fig. 1). This is due to poorer quality reservoirs formed by the conglomerate-rich submarine slope channels or inner fans (Fig. 2d, e). Moreover, the southward tilt of Area 3 due to the post-Pannonian inversion is regarded as a significant seal risk. Nevertheless, both areas are important exploration targets because of the largely untested stratigraphic traps. These are represented by toplap and onlaps of on salt-tectonics related unconformities and lowstand channels incised on the slope.

The eastern parts of the basin are mostly overlain by the Neogene volcanics (Area 4). A characteristic structural feature of this area is the northwest trending diapir lineament (Fig. 1, Eastern Diapir Lineament). Sediments are tilted eastwards and seal risk is high in places not overlain by volcanics. The reservoirs are conglomerate-rich slope channels in the Upper Badenian and sand-rich shallow-marine sandstones in the Sarmatian and Pannonian. As the facies shallows eastwards, the source and seal quality drop rapidly, combined with high amounts of carbon dioxide and nitrogen due to volcanic activity. Thus, hydrocarbon system risk is high. However, the type of depositional systems and structural evolution advocate for largely unexplored stratigraphic traps and various sub-volcanic plays. Structural closures with rollover anticlines are suggested in areas with rugged topography (e.g. Krézsek and Bally, 2006). Stratigraphic pinch-outs on diapir flanks are widespread along the Eastern Diapir Lineament. Among others, the key factor in developing this salt flank play is to understand the effect of post-Pannonian inversion of the diapirs.

Area 5 covers most of the compressional diapirs with shallow décollements (western diapir lineament, Fig. 12) and is considered to have significant hydrocarbon system risk due to limited source thickness (Upper Badenian) and quality (Sarmatian), and the overall lack of reservoirs (commonly thin-bedded turbidites). Stratigraphic pinch-outs have been tested successfully, formed by forced regressive sedimentary wedges encased in mud (e.g. LST5, LST6). Further exploration potential lies in mapping localized sands, e.g. predicting the development of the forced regressive wedges. 4-way dip closures may have a commercial value if better methods of developing thin-bedded sands would be implemented.

For a variety of reasons other areas are not considered very attractive target for exploration. In Area 6 reservoirs are shallow-marine Sarmatian sandstones, subject to large facies changes and the seal is not everywhere present. Some structural closures exist and small stratigraphic closures are expected. The Upper Badenian source is likely missing (or very thin) and the Sarmatian rather uncertain. Area 7 may have better reservoirs than Area 8, but the source and risk is similar. Moreover, the gas-bearing deposits are thin due to erosion and highly tilted. Area 8 has good reservoirs and source, but once more tilted and thus lacks top seals and structural closures. Significant

stratigraphic traps are unlikely because of the sheet-like development of the reservoir sands (mid to outer submarine fan facies associations).

## 9. Conclusions

The most common play type of the Transylvanian Basin is the deep-marine turbidite sandstones sealed by shales in 4-way dip closures related to compressional salt-cored folds. This is a highly explored play in the basin and in our opinion, only limited potential is left.

Our combined study of outcrops, wells and seismic data indicate that several under-explored or new play types exists in this mature basin. In general, these plays have some sort of stratigraphic trapping component (Fig. 12). We highlight the plays formed by onlaps of submarine fans on salt-cored folds and slope channels systems. Such deposits crop out on the basin margins and may be used as reservoir analogues (e.g. Tău, Archita, Coșșa Mică, Cenade, Jimbor, Dacia, etc.; Section 8.2). The best reservoirs are formed by packages of turbidites up to 10 m thick, with mean porosity values of 15–30% and permeabilities of 20–1000 mD. These form tens of meter thick packages especially in submarine fans deposited during sea-level lowstands (Fig. 2a; Krézsek and Filipescu, 2005). The mid fan facies association is typically formed by laterally extensive sheet-like sandstone and shale beds up to 10 m thick with limited vertical connectivity. In turn, inner fans have thicker sandstones (up to 20–30 m) with good vertical connectivity, but are less laterally extensive (<500 m) and may contain conglomerates that decrease reservoir quality.

We attempted to highlight some of the exploration risks related to various play types (Fig. 1). This shows that the central parts of the basin rank with the lowest risk. This is because reservoirs are deep-marine or deep-lacustrine turbidites with minor facies changes sealed by thick shales (Section 4.3). The high risks on the basin margins are due to the late tilt and erosion, the large-scale facies variations in shallow-marine or continental reservoirs and lack of adequate seals (Sections 4.1 and 4.2). Our assessment can be much improved by acquiring more and modern data to address the properties of various petroleum system elements.

## Acknowledgements

This paper is a contribution to the research project ISES 2 Netherlands “Sequence stratigraphy constraints on vertical movements of the Transylvanian Basin border mountains”. We are grateful to Laurențiu Ionescu for comments on earlier versions of the manuscript. Romgaz and the Romanian National Agency for Mineral Resources (ANRM) are thanked for permission to use and publish proprietary data. The authors wish to thank Orsolya Sztanó (ELTE University, Budapest), Nick Culshaw (Dalhousie University, Halifax) and an anonymous reviewer for their reviews and useful comments, which significantly improved the manuscript.

## References

- Andrescu, M., Nielsen, S.B., Polonic, G., Demetrescu, C., 2002. Thermal budget of the Transylvanian lithosphere. Reasons for a low surface heat-flux anomaly in a Neogene intra-Carpathian basin. *Geophysical Journal International* 150, 494–505.
- Baciu, C., Caracausi, A., Etiopie, G., Italiano, F., 2007. Mud volcanoes and methane seeps in Romania: main features and gas flux. *Annals of Geophysics* 50 (4), 501–511.
- Bada, G., Horváth, F., Cloetingh, S., Coblenz, D.D., Tóth, T., 2001. The role of topography induced gravitational stresses in basin inversion: the case study of the Pannonian basin. *Tectonics* 20, 343–363.
- Bertotti, G., Matenco, L., Cloetingh, S., 2003. Vertical movements in and around the south-east Carpathian foredeep: lithospheric memory and stress field control. *Terra Nova* 15 (5), 299–305.
- Blair, T.C., McPherson, J.G., 1994. Alluvial fans and their natural distinction from rivers based on morphology, hydraulic processes, sedimentary processes, and facies assemblages. *Journal of Sediment Research* A64, 450–489.
- Böckh, H., 1911. Az erdélyi medence földgázellőfordulásainak geológiája. *Bányászati és Kohászati Lapok XLIV (II)*, 75. Budapest.

- Bouma, A.H., 1962. Sedimentology of Some Flysch Deposits, a Graphic Approach to Facies Interpretation. Elsevier, Amsterdam, 168 pp.
- de Broucker, G., Mellin, A., Duindam, P., 1998. Tectono-Stratigraphic evolution of the Transylvanian Basin, pre-salt sequence, Romania. In: Dinu, C., Mocanu, V. (Eds.), Geological and Hydrocarbon Potential of the Romanian Areas. Bucharest Geosciences Forum, vol. 1, pp. 36–70.
- Bucur, I.L., Piteiu, M.-A., Săsăran, E., 2004. The Mesozoic carbonate deposits from the borehole 6042 Deleni (Transylvanian Depression), Studia UBB. Geologia 59/2, 27–48.
- Ciulavu, D., Dinu, C., Szakács, A., Dordea, D., 2000. Neogene kinematics of the Transylvanian Basin, Romania. AAPG Bulletin 84, 1589–1615.
- Ciupagea, D., Păucă, M., Ichim, T., 1970. Geology of the Transylvanian depression. Academiei Republicii Socialiste Romania, Bucharest, 256 (in Romanian).
- Clayton, C., 1992. Source volumetrics of biogenic gas generation. In: Vially, R. (Ed.), Bacterial Gas. Editions Technip, Paris, pp. 191–204.
- Cloetingh, S., Burov, E., Mañenco, L., Toussaint, G., Bertotti, G., Andriessen, P.A.M., Wortel, M.J.R., Spakman, W., 2004. Thermo-mechanical controls on the mode of continental collision in the SE Carpathians (Romania). Earth and Planetary Science Letters 218 (1–2), 57–76.
- Cloetingh, S., Bada, G., Mañenco, L., Lankreijer, A., Horváth, F., Dinu, C., 2006. Models of basin (de)formation, lithospheric strength and vertical motions in the Pannonian-Carpathian system: inferences from thermo-mechanical modeling. In: Gee, D.G., Stephenson, R.A. (Eds.), European Lithosphere Dynamics. Geological Society, London, Memoirs 32, pp. 207–221.
- Crânganu, C., Deming, D., 1996. Heat flow and hydrocarbon generation in the Transylvanian Basin, Romania. AAPG Bulletin 10, 1641–1653.
- Csontos, L., Nagymarosy, A., 1998. The Mid-Hungarian line: a zone of repeated tectonic inversions. Tectonophysics 297, 51–71.
- Csontos, L., Vörös, A., 2004. Mesozoic plate reconstructions of the Carpathian region. Palaeogeography, Palaeoclimatology, Palaeoecology 210, 1–56.
- Demetrescu, C., Veliciu, S., Burst, A.D., 1992. Heat flow map of Romania. In: Hurlig, E., Cermák, V., Haenel, R., Zui, V. (Eds.), Geothermal Atlas of Europe. VEB Hermann Haack, Gotha, pp. 72–74.
- Falk, P.D., Dorsey, R.J., 1998. Rapid development of gravelly high-density turbidity currents in marine Gilbert type fan deltas, Loreto Basin, Baja California Sur, Mexico. Sedimentology 45, 331–350.
- Filipescu, S., 1992. Data concerning the Kossovian from Podeni (western border of the Transylvanian Basin). Studia Univ. Babeş-Bolyai. Geologia XXXVII (2), 79–82.
- Filipescu, S., 1999. The significance of foraminifera fauna from the Feleac Formation (Transylvanian Basin, Romania). Studia Univ. Babeş-Bolyai. Geologia XLIV (2), 125–131. Cluj-Napoca.
- Filipescu, S., 2004a. Agglutinated foraminifera to sequence stratigraphy – the case of Bogdanowiczia pocutia Pishvanova. In: Bubik, M., Kaminski, M.A. (Eds.), Proceedings of the Fifth International Workshop on Agglutinated Foraminifera. Grzybowski Foundation Special Publication, vol. 8, p. 460.
- Filipescu, S., 2004b. Anomalinoidea dividens bioevent at the Badenian/Sarmatian boundary – a response to paleogeographic and paleoenvironmental changes. Studia Universitatis Babeş-Bolyai. Seria Geologia 69 (2), 21–26.
- Filipescu, S., Silye, L., Krézsek, C., 2005. Sarmatian micropaleontological assemblages and sedimentary paleoenvironments in the Southern Transylvanian Basin. Acta Palaeontologica Romaniaica 5, 173–179.
- Filipescu, S., Silye, L., 2008. New biozones of planktonic foraminifera in the Middle Miocene of the Transylvanian Basin as potential opportunities for the Paratethyan stratigraphy. Geologica Carpathica 59 (6), 537–544.
- Fodor, L., Csontos, L., Bada, G., Györfi, I., Benkovic, L., 1999. Tertiary tectonic evolution of the Pannonian Basin system and neighbouring orogens: a new synthesis of paleostress data. In: Durand, B., Jolivet, L., Horváth, F., Séranne, M. (Eds.), The Mediterranean Basins: Tertiary Extension within the Alpine Orogen. Geological Society, Special Publication, vol. 156, pp. 295–334.
- Fügenschuh, B., Schmid, S.M., 2005. Age and significance of core complex formation in a very curved orogen: evidence from fission track studies in the South Carpathians (Romania). Tectonophysics 404, 33–53.
- Gheorghian, M., 1975. Faziostatotipus: Tălmăcel, dans le sud de la Transylvanie (Roumanie). In: Báldi, T., Senes, J. (Eds.), Chronostratigraphie und Neostartotypen, Miozän der Zentralen Paratethys V, OM Egerin, pp. 176–178. Bratislava.
- Gheorghian, M., Gheorghian, M., Iva, M., 1967. Microfauna and Paleogeography of the Tortonian near Sibiu (Transylvanian Basin), vol. 43/1. Darea de Seamă Institute of Geology, pp. 329–340 (in Romanian).
- Ghergari, L., Mészáros, N., Hosu, A., Filipescu, S., Chira, C., 1991. The gypsiferous formation at Cheia (Cluj County). Studia Universitatis Babeş-Bolyai. Seria Geologia 36 (1), 13–28.
- Gradstein, F.M., Ogg, J.G., et al., 2004. A Geologic Time Scale 2004. Cambridge University Press, 589 pp.
- Gröger, H.R., 2006. Thermal and structural evolution of the East Carpathians in northern Romania: from Cretaceous orogeny to final exhumation during Miocene collision. PhD thesis, Basel University, Basel, 111 pp.
- Harzhauser, M., Piller, W.E., 2007. Benchmark data of a changing sea – Palaeogeography, Palaeobiogeography and events in the Central Paratethys during the Miocene. Palaeogeography, Palaeoclimatology, Palaeoecology 253, 8–31.
- Helland-Hansen, D., Siggerud, E., Bell, B., Benn, J., Dennehy, C., Herman, O., 2007. Rosebank: oil field appraisal/development in the volcanic province of the north west Atlantic margin. Abstract. In: AAPG & AAPG European Region Energy Conference and Exhibition, November 18–21, 2007, Athens, Greece.
- Horváth, F., 1993. Towards a mechanical model for the formation of the Pannonian basin. Tectonophysics 226, 333–357.
- Hosu, A., Filipescu, S., 1996. Sequence stratigraphy on the Middle-Miocene deposits from Gârbova de Sus (Transylvanian Basin, Romania). Studii și Comunicări, Muzeul Județean Bistrița 1, 87–97.
- Huismans, R.S., Bertotti, G., 2002. The Transylvanian Basin, transfer zone between coeval extending and contracting regions. Inferences on the relative importance of slab pull and rift push in arc – back-arc systems. Tectonics 21 (2), 1–19.
- Huismans, R.S., Bertotti, G., Ciulavu, D., Sanders, C.A.E., Cloetingh, S., Dinu, C., 1997. Structural evolution of the Transylvanian Basin (Romania): a sedimentary basin in the bend of the Carpathians. Tectonophysics 272, 249–268.
- Ilie, M.D., 1955. Cercetări geologice în Bazinul Transilvaniei (regiunea Alba-Iulia–Sibiu–Făgărași–Rupea), vol. 28. Anuarul Comitetului Geologic, pp. 255–361.
- Ismail-Zadeh, A., Mueller, B., Schubert, G., 2005. Three-dimensional numerical modeling of contemporary mantle flow and tectonic stress beneath the earthquake-prone southeastern Carpathians based on integrated analysis of seismic, heat flow, and gravity data. Physics of the Earth and Planetary Interiors 149, 81–98.
- Kotarba, M.J., Wilczek, T., Kosakowski, P., Kowalski, A., Wieclaw, D., 1998. A study of organic matter and habitat of gaseous hydrocarbons in the Miocene strata of the Polish part of the Carpathian Foredeep. Przegląd Geologiczny 46, 742–750.
- Kovács, J.Sz., Piteiu, M., Bucur, I.L., Săsăran, E., Pinca, O.G., Nagy, I., 2007. Carbonate reservoirs of the 6042 Deleni well (Târnava Basin, Romania): questions arising for future exploration. Abstract. In: AAPG & AAPG European Region Energy Conference and Exhibition, November 18–21, 2007, Athens, Greece.
- Kovač, M., Holcova, K., Nagymarosy, A., 1999. Paleogeography, paleobathymetry and relative sea-level changes in the Danube Basin and adjacent areas. Geologica Carpathica 50, 325–338.
- Krészsek, C., Bally, A.W., 2006. The Transylvanian Basin (Romania) and its relation to the Carpathian fold and thrust belt: insights in gravitational salt tectonics. Marine and Petroleum Geology 23, 405–442.
- Krészsek, C., Filipescu, S., 2005. Middle to Late Miocene sequence stratigraphy of the Transylvanian Basin (Romania). Tectonophysics 410 (1–4), 437–463.
- Krzywiec, P., 2001. Contrasting tectonic and sedimentary history of the central and eastern parts of the Polish foredeep basin – results of seismic data interpretation. Marine and Petroleum Geology 18, 13–38.
- Lowe, D.R., 1982. Sediment gravity flows: II. Depositional models with special reference to the deposits of high-density turbidity currents. Journal of Sedimentary Petrology 52, 279–297.
- Lubenscu, V., 1981. Biostratigraphic Study on the Upper Neogene from the South-western Transylvania, vol. 58. Anuarul Institutului de Geologie și Geofizică București, pp. 123–194 (in Romanian).
- Magyar, I., Geary, D.H., Müller, P., 1999. Paleogeographic evolution of the Late Miocene Lake Nannon in Central Europe. Palaeogeography, Palaeoclimatology, Palaeoecology 147, 151–167.
- Mârza, I., Mészáros, M., 1991. Les tuffs volcaniques de Transylvanie: historique, valeur theorique et pratique dans le developpement de la geologie Transylvainie. In: Bedelean, I., Ghergari, L., Mârza, I., Mészáros, N., Nicorici, E., Petrescu, I. (Eds.), The Volcanic Tuffs from the Transylvanian Basin, pp. 11–21. Romania, Cluj Napoca.
- Mason, P.R.D., Seghedi, I., Szakacs, A., Downes, H., 1998. Magmatic constraints on geodynamic models of subduction in the East Carpathians, Romania. Tectonophysics 297, 157–176.
- Mañenco, L., Bertotti, G., Leever, K., Cloetingh, S., Schmid, S.M., Tărăpoancă, M., Dinu, C., 2007. Large-scale deformation in a locked collisional boundary: interplay between subsidence and uplift, intraplate stress, and inherited lithospheric structure in the late stage of the SE Carpathians evolution. Tectonics 26, 1–29.
- Mañenco, L., Schmid, S.M., 1999. Exhumation of the Danubian nappes system (South Carpathians) during the Early Tertiary: inferences from kinematic and paleostress analysis at the Getic/Danubian nappes contact. Tectonophysics 314, 401–422.
- Miall, A.D., 1996. The Geology of Fluvial Deposits. Sedimentary Facies, Basin Analysis and Petroleum Geology. Springer-Verlag, Berlin, 582 pp.
- Morariu, D.C., 1998. Subtle traps in the petroleum hydrocarbon systems of Romania. Ph.D. thesis, Bucharest University, Romania, 156 pp.
- Mrázek, L., Jekelius, E., 1927. Aperçu sur la structure du bassin neogene de Transylvanie. Guide des Excursions, deuxième Reunion Associations pour l'Avancement de la Geologie des Carpatés, Bucharest.
- Nemeșan, M., 2007. Romgaz. Present and Perspectives. Energy in Center and Eastern Europe Forum, Bucharest, November 20–22, 2007.
- Paraschiv, D., 1979. Romanian Oil and Gas Fields, vol. 13. Institutul de Geologie și Geofizică Studii Tehnice și Economice Seria A, Bucharest, 382 pp.
- Paraschiv, D., 1980. Observații asupra unor zăcăminte de gaze mixte din partea de Est a Depresiunii Transilvaniei. Seria Geologia 25, 83–93. St. Cerc. Geol. Geofiz. Geogr.
- Paraschiv, D., 1997. The pre-Paratethys buried denudational surface in Romanian territory. Revue Roumaine de Geographie 41, 21–32.
- Pawlewitz, M., 2005. Transylvanian Composite Total Petroleum System of the Transylvanian Basin Province, Romania, Eastern Europe. USGS Bulletin 2204-E, 10 p.
- Pécskay, Z., Lexa, J., Szakács, A., Balogh, K., Seghedi, I., Konecny, V., Kovács, M., Márton, E., Kaliciak, M., Székely-Fux, V., Póka, T., Gyarmati, P., Edelstein, O., Roșu, E., Zec, B., 1995. Space and time evolution of the Neogene–Quaternary volcanism in the Carpatho-Pannonian Region. Acta Vulcanologica 7 (2), 15–28.
- Peryt, T.M., 2006. The beginning, development and termination of the Middle Miocene Badenian salinity crisis in Central Paratethys. Sedimentary Geology 188–189, 379–396.
- Petrescu, I., Mészáros, N., Filipescu, S., Buda, A., 1988. Contributions to the stratigraphic knowledge of the Neogene deposits in Borehole 12 Merești (SE Transylvanian Basin). Studia Universitatis Babeş-Bolyai. Seria Geologia-Geographia 33/2, 15–24.

- Popescu, B.M., 1995. Romania's petroleum systems and their remaining potential. *Petroleum Geoscience* 1, 337–350.
- Popescu, G., Mărunțeanu, M., Filipescu, S., 1995. Neogene from Transylvanian depression. Guide to excursion A1 Xth RCMNS Congress București, 4–9 September 1995. *Romanian Journal of Stratigraphy* 76 (3), 27.
- Postma, G., 1990. Depositional Architecture and Facies of River and Fan Deltas: a Synthesis, vol. 10. IAS Special Publication, pp. 13–27.
- Reading, H., Richards, M., 1994. The classification of deep-water siliciclastic depositional systems by grain size and feeder systems. *AAPG Bulletin* 78, 792–822.
- Richards, M., Bowman, M., 1998. Submarine fans and related depositional systems II: variability in reservoir architecture and wireline log character. *Marine and Petroleum Geology* 15, 821–839.
- Rögl, F., 1996. Stratigraphic correlation of the Paratethys Oligocene and Miocene. In: Decker, K. (Ed.), *Mitteilungen der gesellschaft der geologie und bergbaustudenten in Osterreich*, vol. 41, pp. 65–73.
- Roure, F., Roca, E., Sassi, W., 1993. The Neogene evolution of the outer Carpathian flysch units (Poland, Ukraine and Romania): kinematics of a foreland/ fold-and-thrust belt system. *Sedimentary Geology* 86, 177–201.
- Royden, L., 1988. Late Cenozoic tectonics of the Pannonian Basin system, *Memoir* 45. AAPG, 27–48.
- Răbăgia, T., Mațenco, L., 1999. Tertiary tectonic and sedimentological evolution of the South Carpathians foredeep: tectonic versus eustatic control. *Marine and Petroleum Geology* 16, 719–740.
- Rădulescu, F., 1988. Seismic models of the crustal structure in Romania. *Revue Roumaine de Geologie, Geophysique et Geographie, Serie de Geologie* 32, 13–17.
- Sachsenhofer, R.F., 1996. The Neogene Styrian Basin: an Overview, vol. 41. *Mitteilungen der Gesellschaft der Geologie- und Bergbaustudenten in Österreich*, 19–32.
- Saint Martin, J.-P., Merle, D., Cornée, J.-J., Filipescu, S., Saint Martin, S., Bucur, I.I., 2007. Les constructions coralliennes du Badénien (Miocène moyen) de la bordure occidentale de la dépression de Transylvanie (Roumanie). *Comptes Rendus Palevol* 6, 37–46.
- Sanders, C.A.E., Andriessen, P., Cloetingh, S., 1999. Life cycle of the East Carpathian orogen: erosion history of a doubly vergent critical wedge assessed by fission track thermochronology. *Journal of Geophysical Research* 104 (B12), 29095–29112.
- Sanders, C.A.E., Huisman, R., van Wees, J.D., Andriessen, P., 2002. The Neogene history of the Transylvanian Basin in relation to its surrounding mountains. *EGU Stephan Mueller Special Publication Series* 3, 121–133.
- Schmid, S.M., Berza, T., Diaconescu, V., Frotzheim, N., Fügenschuh, B., 1998. Orogen-parallel extension in the South Carpathians during the Paleogene. *Tectonophysics* 297, 209–228.
- Schmid, S.M., Bernoulli, D., Fügenschuh, B., Mațenco, L., Schefer, S., Schuster, R., Tischler, M., Ustaszewski, K., 2008. The Alpine–Carpathian–Dinaridic orogenic system: correlation and evolution of tectonic units. *Swiss Journal of Geosciences* 101 (1), 139–183.
- Schuller, V., 2004. Evolution and geodynamic significance of the Upper Gosau basin in the Apuseni Mountains (Romania). PhD thesis, *Tübinger Geowiss. Arb.*, A70, 112 pp.
- Schultz, H.-M., van Berk, W., Bechtel, A., Struck, U., 2009. Bacterial methane in the Atzbach–Schwanenstadt gas field (Upper Austrian Molasse Basin). Part I: geology. *Marine and Petroleum Geology* 26, 1180–1189.
- Steininger, F., Senes, J., 1971. M1 Eggenburgien. Die Eggenburger Schichten-gruppe und ihr Stratotypus, vol. 2. *Chronostratigraphie und Neostatotypen*, pp. 1–827.
- Surlyk, F., 1984. Fan-delta to submarine fan conglomerates of the Volgian–Valanginian Wollaston Forland Group, East Greenland. In: Koster, E.H., Steel, R.J. (Eds.), *Sedimentology of Gravels and Conglomerates*. Canadian Society of Petroleum Geologists, *Memoir* 10, pp. 359–382.
- Szakács, A., 2003. Mineral chemistry of the primary magmatic mineral assemblage of the “Dej Tuff” (Romania). *Studia Universitatis “Babeș-Bolyai”. Seria Geologia Special issue V*, 116–120.
- Szakács, A., Seghedi, I., 1995. The Călimani–Gurghiu–harghita volcanic chain, East Carpathians, Romania: volcanological features. *Acta Vulcanologica* 7 (2), 145–153.
- Sztanó, O., Krézsek, C., Magyar, I., Wanek, F., Juhász, G., 2005. Sedimentary cycles and rhythms in a Sarmatian to Pannonian (Late Miocene) transitional section at Oarba de Mureș/Marosorbó, Transylvanian Basin. *Acta Geologica Hungarica* 48, 235–257.
- Săndulescu, M., 1988. Cenozoic tectonic history of the Carpathians. In: Royden, L., Horváth, F. (Eds.), *The Pannonian Basin: a Study in Basin Evolution*, *Memoir* 45. AAPG, pp. 17–25.
- Tari, G.C., 1994. Alpine tectonics of the Pannonian basin. PhD thesis, Rice University, Houston, 501 pp.
- Tari, G.C., Dövényi, P., Dunkl, I., Horváth, F., Lenkey, L., Ștefănescu, M., Szafián, P., Tóth, T., 1999. Lithospheric structure of the Pannonian basin derived from seismic, gravity and geothermal data. In: Durand, B., Jolivet, L., Horváth, F., Séranne, M. (Eds.), *The Mediterranean Basins: Tertiary Extension within the Alpine Orogen*. Geological Society, Special Publication, vol. 156, pp. 215–250.
- Tari, G.C., Horváth, F., 2006. Alpine evolution and hydrocarbon geology of the Pannonian Basin: an overview. In: Golonka, J., Picha, F. (Eds.), *The Carpathians and their Foreland: Geology and Hydrocarbon Resources*, *Memoir* 84. AAPG, pp. 605–618.
- Tiliță, M., Lenkey, L., Mațenco, L., Horváth, F., Dinu, C., 2006. Neogene evolution of Transylvania basin: insights derived from (2D steady-state) thermal modeling. *Geophysical Research Abstracts* 8, 08874. SRef-ID: 1607-7962/gra/EGU06-A-08874.
- Tischler, M., Mațenco, L., Filipescu, S., Gröger, H.R., Wetzel, A., Fügenschuh, B., 2008. Tectonics and sedimentation during convergence of the ALCAPA and Tisza–Dacia continental blocks: the Pienide nappe emplacement and its foredeep (N. Romania). In: Siegesmund, S., Fügenschuh, B., Frotzheim, N. (Eds.), *Tectonic Aspects of the Alpine–Dinaride–Carpathian System*. Geological Society, London, Special Publications, vol. 298, pp. 317–334.
- Vancea, A., 1960. The Neogene in the Transylvanian Basin. *Academiei Republicii Socialiste Romania, Bucharest*, 262 (in Romanian).
- Vendeville, B.C., Jackson, M.P.A., 1992. The rise of diapirs during thin-skinned extension. *Marine and Petroleum Geology* 9, 331–353.
- Visarion, M., Veliciu, S., 1981. Some considerations on the low heat flow in the Transylvanian basin. *Studii și Cercetari de Geologie, Geofizica, Geografie-Geofizica* 19, 53–60 (in Romanian).
- Wanek, F., 1999. Exploration for mineral waters and hydrocarbons in the Eastern Carpathians before 1908. *Acta 1998/1. Székely Nemzeti Múzeum*, pp. 45–56 (in Hungarian).
- Wortel, M.J.R., Spakman, W., 2000. Subduction and slab detachment in the Mediterranean–Carpathian region. *Science* 290, 1910–1917.