The Transylvanian Basin (Romania) and its Relation to the Carpathian Fold and Thrust Belt: Insights in Gravitational Salt Tectonics

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ABSTRACT
Interpretation of regional seismic profiles, stratigraphic and sedimentologic data improved insights in the evolution of the Transylvanian Basin. The basin evolution was coeval with the post Mid-Cretaceous to recent deformation of the Carpathian Mts. Four tectonostratigraphic megasequences are differentiated: Upper Cretaceous (rift), Paleogene (sag), Lower Miocene (flexural basin) and Middle to Upper Miocene (backarc sequence dominated by gravitational tectonics). The Mid-Miocene continental collision in the Eastern Carpathians is associated with the rising Carpathians. This uplift enhanced the differential load, which together with the high heat flow induced by Late Miocene to Pliocene arc volcanism, triggered large-scale Mio-Pliocene gravity spreading of the salt overburden. This "mega-slide" comprises three structural domains, as follows: extensional weld (upslope), contractional folds (central) and contractional toe thrust (downslope). The diapirs in the east indicate a pre-shortening reactive/passive growth stage. The central folds are mostly the result of late shortening. Basement involved thrusting uplifted the toe thrust domain by the Late Pliocene. The Late Neogene to recent Carpathians uplift, backarc volcanism and gravity spreading are largely coeval.

KEYWORDS: Transylvanian Basin; Carpathians; salt tectonics.

INTRODUCTION

The Carpathians and the Dinarides are the eastward continuation of the European Alpine thrust and fold belts. These formed during the collision of the African promontory with Europe (e.g. Bally and Snelson, 1980; Ziegler, 1988; Tari et al., 1995). During the Late Cretaceous and Cenozoic, various basin systems formed on the top of the "Eo-Alpine" structures, such as those situated in the intra-Carpathian area (i.e. the present relatively lowland area bordered by the Eastern Alps, Carpathians and Dinarides) (Fig. 1).

The Carpathians were traditionally divided into two major units (e.g. Uhlig, 1907): the Internal and External Carpathians. The Internal Eastern Carpathians were formed in Mid-Cretaceous time and involve Precambrian and/or Paleozoic crystalline basement and its Mesozoic (pre-Cenomanian) sedimentary cover (e.g. Săndulescu, 1988). Upper Cretaceous to Paleogene post-tectonic sedimentary formations (wedge-top basin fills) locally cover the nappes in the Transylvanian Basin.

The External Eastern Carpathian thrust and fold belt is a stack of eastward-verging thin-skinned imbricates that rode on top of the subducted East European plate and formed during the Mid-Cretaceous to Pleistocene. This "accretionary wedge" now overrides the relatively undeformed Eastern European foreland. Except for the most external, and youngest, mollasse units, the nappes comprise only flysch-type deposits of Cretaceous to Lower Miocene age.

The Transylvanian Basin represents a post-Cenomanian sedimentary basin developed on top of the Middle Cretaceous basement nappes (internal Carpathians). The nappes form the hinterland of the Carpathians "backstop" of the foreland folded belt further to the east. The Mid to Late Miocene Transylvanian Basin, among other intra-Carpathian basins (e.g. Royden, 1985; Tari and Horváth, 1995; Kováč et al. 1995; Decker, 1996; Sachsenhofer, 1996) was situated behind the Carpathian arc (Fig. 1, 6). While most of these basins had a typical backarc-type evolution (syn-rift extension leading to thinning of the continental crust, followed by post-rift thermal subsidence) (e.g. Bally and Snelson, 1980; Royden, 1988; Csontos et al., 1992; Tari and Horváth, 1995; Bada, 1999; Fodor et al., 1999), the Transylvanian Basin tectonic and sedimentary history is markedly different (Krézsek and Filipescu, 2005; Tiliță et al., 2006).

These differences are best illustrated by some geophysical characteristics. The Transylvanian Basin has a relatively thick continental crust (Visarion and Veliciu, 1981; Răileanu et al., 1994) and regional low surface heat flow (e.g. Demetrescu et al., 2001). This contrasts with the thinned continental crust (e.g. Tari et al., 1999) and high heat flow of the Pannonian Basin (e.g. Dövényi and Horváth, 1988; Fig. 1). Other striking features of the Transylvanian Basin are the lack of Miocene extensional structures, except for superficial extension related to salt tectonics (e.g. De Broucker et al., 1998; Krézsek and Filipescu, 2005) and the very high subsidence rates during the Mid to Late Miocene (e.g. Crânganu and Demming, 1996). One of the most important factors influencing the late stage of the Transylvanian Basin

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is related to the Late Miocene uplift of the Carpathians (e.g. Sanders, 1999) and associated gravitational spreading (cf. Shultz-Ela, 2001) of the post-salt overburden strata (Krézsek, 2004).

Several papers have dealt with the tectonic evolution of the Eastern Carpathians (e.g. Săndulescu, 1988; Gîrbacea and Frisch, 1998; Zweigel et al., 1998; Linzer et al., 1998; Sanders, 1999; Maţenco and Bertotti, 2000; Gibson, 2001; Cloetingh et al., 2004). However, only a few discuss the tectonic/sedimentary relationship with the foreland (Răbăgia and Maţenco, 1999; Maţenco et al. 2003; Tărăpoancă et al., 2003, 2004) or hinterland (De Broucker et al., 1998; Ciulavu et al., 2000; Sanders et al., 2002; Krézsek and Filipescu, 2005).

The Transylvanian Basin and the Carpathians are where the scientific study of salt tectonics became established during the late 19th to early 20th century (Jackson, 1997). Pošepný (1871) first illustrated a salt dome from Transylvanian and the term "diapir" was proposed by Mrazec (1907) for salt-structures found in the Carpathians folded and thrust belt (Ştefănescu et al., 2000).

Our goal here is to present the evolution of the Transylvanian Basin and discuss its relations to the Carpathian fold belt and its foreland. We will emphasize the role of salt tectonics in the late stage basin development and its relation to the evolution of the Carpathians.

THE BASEMENT

The Transylvanian Basin basement is shared with the Eastern Pannonian Basin (Fig. 1; e.g. Tari et al., 1999; Csontos and Vörös, 2004) and consists of a stack of basement-involved thrust sheets, which were assembled by the Mid Cretaceous. These involve Paleozoic crystalline units (e.g. Săndulescu and Visarion, 1978; Visarion and Veliciu, 1981; Săndulescu, 1988), ophiolites and island-arc volcanics (Fig. 2; e.g. Burchfiel, 1976; Bortolotti et al. 2004 and references therein), and sedimentary cover of Triassic to Mid Cretaceous age (e.g. Ciupagea et al., 1970; Burchfiel, 1976; Paraschiv, 1979). The combined nappes are referred to as the Tisa-Dacia terrane (e.g. Balla, 1987; Csontos et al., 1992, Tari et al., 1995; Fodor et al., 1999; Csontos and Vörös, 2004).

The assembled tectonic units evolved from a Triassic to Lower Cretaceous intercontinental rift or narrow ocean ("Transylvanian Ocean") and its passive margins (Tisa, Dacia) (Fig. 2, 3). By the Mid Cretaceous the "Transylvanian Ocean" was subducted and its former passive margins collided, forming thick-skinned nappes. We will only use the terms Tisa, Transylvanides and Dacia, without discussing any details of their structural evolution. Turonian to Campanian deep-marine sediments deposited in post-collision piggyback/wedge-top basins locally seal the Middle Cretaceous nappe contacts (Fig. 2) (e.g. Burchfiel, 1976; Bleahu et al., 1981; Săndulescu, 1988; Pană and Erdmer, 1994; Dallmayer et al., 1999; Csontos and Vörös, 2004)

The assemblage of Middle Cretaceous basement thrusts formed the "backstop" of the Cenozoic Carpathian foreland fold belt or "accretionary wedge". Thus, we differentiate the following basinal domains (Fig. 2): 1) foredeep basins formed in the front of the Carpathian accretionary wedge; 2) wedge-top basins (piggyback or satellite basins) on the Carpathian accretionary wedge and 3) intra-Carpathian basins formed on top of the backstop (i.e. the pre-Upper Cretaceous stack of basement sheets; the Tisa-Dacia terrane). The Transylvanian Basin is one of these intra-Carpathian basins (Fig. 1).

The remnants of the "Transylvanian Ocean" (Transylvanides), i.e. the "Transylvanian suture" ophiolite and cover outcrop in the South Apuseni Mts. (e.g. Bortolotti et al. 2004), and may be followed in the subsurface, toward the western and central parts of the Transylvanian Basin (Fig. 3) (Burchfiel, 1976 Săndulescu and Visarion, 1978; Visarion and Veliciu, 1981; Stănică et al., 2000). The ophiolites and their cover do not appear to be significantly thickened by subduction-related thrusting in the subsurface of the central Transylvanian Basin. Alternatively, they are relatively thin oceanic successions obducted onto the passive margin of Dacia (e.g. Săndulescu, 1988) or perhaps they were not involved in the subduction (Maţenco et al., 2005). Obducted Transylvanides on Dacia-type basement outcrop in the Inner Eastern Carpathians (Fig. 3; Burchfiel, 1976; Săndulescu, 1988).

Of the more than 4000 wells drilled to explore for hydrocarbons in the Transylvanian Basin, only about 50 penetrated the Transylvanian Basin basement (Fig. 3, Table 1). In addition to crystalline basement and island arc-type volcanics (IAV) of the various allochthonous units, a few wells have penetrated thick sequences of Jurassic and Cretaceous platform carbonates. Correlating these carbonate sequences to the outcrops surrounding the Transylvanian Basin is problematic, particularly for the Deleni (Fig. 4) or Band wells which encountered thick Lower Cretaceous and Jurassic carbonates (Paraschiv, 1979). In the Deleni well, the Upper Jurassic is represented by shallow prograding platform carbonates (from inner ramp to supratidal), followed by Lower Cretaceous outer shelf and Upper Cretaceous deep-marine siliciclastics (Bucur and Săsăran, 2004). This deepening and the coeval change from carbonate to siliciclastic deposition may be related to the onset of Mid Cretaceous subduction of the "Transylvanian Ocean" which, in turn, triggered increased subsidence in the foredeep area of the Transylvanian subduction zone.

UPPERMOST CRETACEOUS AND CENOZOIC STRATIGRAPHY

The Paratethys and standard (Tethys) versus regional (Paratethys) subdivisions

Uplift of the Alpine – Carpathian chain began during the Late Paleogene. This uplift limited water exchanges with the main Tethyan Ocean and thereby triggering the evolution of endemic faunas in the restricted areas (e.g. Rögl, 1996). This restricted oceanic realm once stretched from the Alpine molasse basin to the present Aral Sea, and is known as the Paratethys (Fig. 1) (Laskarev, 1924). Because of the endemic nature of the Paratethyan faunas, a direct correlation with the Tethyan faunas is difficult. The Paratethys may be further divided into aquatic bioprovinces. In general, three Paratethyan realms have been recognized (Western, Central and Eastern; Fig. 1) and local stages were introduced for each in order to describe the stratigraphic evolution of these basins. (e.g. Rögl, 1996; Popov et al., 2004; Gradstein et al., 2004 and references therein). The Transylvanian Basin is part of the Central Paratethys, so we will use the corresponding regional stages (for correlations with the standard stages please refer to Fig. 5).

Tectonostratigraphic megasequences

The sedimentary record of any given basin commonly reflects major tectonic regimes (e.g rifting, uplift, inversion, tilting, flexure, etc.) that have a significant influence on sedimentation and accommodation space. Changes of these tectonic regimes leads to the formation of major regional unconformities that divide the sedimentary record into tectonostratigraphic megasequences (e.g. Hubbard, 1988 cf. Sharland et al., 2001). They usually develop during a time-span of more than 10 My and often coincide with the second-order cycles of sequence stratigraphy (e.g. Vail et al., 1977).

The stratigraphic section in the Transylvanian Basin is locally more than 5 km thick and may be divided into four tectonostratigraphic megasequences (Fig. 5): 1) Upper Cretaceous (rift, gravitational collapse), 2) Paleogene (sag), 3) Lower Miocene (flexural), and 4) Middle to Upper Miocene (backarc sequence dominated by gravitational tectonics). Several papers describe the bio- and litho-stratigraphy of these megasequences (e.g. Koch, 1894, 1900; Vancea, 1960; Ciupagea et al., 1970; Dicea et al., 1980a; Petrescu et al., 1987, 1989; Rusu, 1989, 1995; Filipescu, 1996, 2001; Mészáros, 2000) and a few deal with Eocene and Miocene 3rd order sequence stratigraphy (Proust and Hosu, 1996; Krézsek and Filipescu, 2005). This paper will confine itself to the discussion of the tectonostratigraphic megasequences.

General Lithostratigraphy

Upper Cretaceous (Santonian – Maastrichtian) sediments that seal the basement nappes consist of conglomerates, sandstones, marls and rudist limestones deposited in continental, shallow-marine and deep-marine settings (e.g. Paraschiv, 1979; De Broucker et al., 1998). The succession has variable thickness (few 100's to more then 1000 m). Similar deposits outcrop in the Apuseni Mountains and the Carpathians (e.g. Bleahu et al., 1981; Lupu et al., 1993; Willingshofer et al., 1999; Schuller, 2004; and references therein).

The Paleocene and Lower Eocene are dominated by continental (Jibou Formation) conglomerates, sandstones and shales with thickness ranging from a few to more than 500 meters (e.g. Koch, 1894; Proust and Hosu, 1996). The Middle to Upper Eocene is represented by the Călata and Turea Groups and includes limestones, sandstones, marls and evaporites (e.g. Rusu, 1995; Filipescu, 2001, and references therein). They were deposited in a shelf to deep-marine setting. The Oligocene represented by mainly siliciclastic sediments and consists of conglomerates, sandstones, coals and bituminous shales. In general the shallow-marine Eocene – Oligocene succession is a few hundred meters thick, while the deeper marine section is more than 500-1000 m thick (e.g. Paraschiv, 1979).

The Lower Miocene is represented by shallow-marine sandstones (Coruş Formation), outer shelf marls (Chechiş Formation), and coarse-grained fan deltas and deep-marine deposits (Hida Formation) (e.g. Popescu et al., 1995; Filipescu, 2001 and references therein). In the southern part of the basin, continental and shallow-marine deposition occurred locally. In general, the thickness of the Lower Miocene succession decreases from more than 1000 m in the north to zero in the south.

The lower part of the Middle Miocene is composed of coarse-grained fan delta conglomerates, deep-marine deposits with thick tuff horizons (Dej Formation) and local thin (<100 m) shallow-marine carbonates (Gârbova Formation) (Filipescu, 2001, and references therein). Regionally important evaporite deposition occurs in the Mid Badenian, consisting of gypsum (basin margins) and salt (basin center). They were deposited in a deep, but desiccated basin (Krézsek and Filipescu, 2005). The depositional thickness of the salt may have been around 300 m. The post-salt succession is dominated by deep-marine and outer ramp siliciclastic deposits. Shallow-water (shelf and delta) conglomerates, sandstones and shales became important in the Late Sarmatian (Krézsek and Filipescu, 2005). The post-salt deposits thicken from the NW (a few hundred meters) to the SE (more than 2000 m).

The youngest preserved sedimentary fill of the Transylvanian Basin is represented by Upper Miocene sandstones, marls and conglomerates (Krézsek, 2005). Most of the Upper Miocene fill of the basin was eroded due to Late Badenian to Pliocene SE-ward tilting (De Broucker et al., 1998; Krézsek and Filipescu, 2005), and regional Pliocene uplift and erosion of the Eastern Carpathians (Sanders, 1999).

During the Pliocene, minor basins (Braşov area and north of it) formed on top of the Carpathians. Some authors think that these basins were formed by extension related to slab detachment and delamination (Gîrbacea and Frisch, 1998; Chalot-Prat and Gîrbacea, 2000) or by lithospheric buckling (Bertotti et al., 2003). Paleontological evidence (Wanek,

2004 pers. com.) and fission-track data (Sanders, 1999) suggest that the Pliocene basins had connections with the Carpathian foredeep at least through the present-day Carpathian bend area, i.e. where the orogen changes orientation by 80 degrees from the nearly N-S striking Eastern Carpathians into the E-W striking Southern Carpathians (e.g. Gibson, 2001; Fig. 6). This suggests that the Carpathian bend was not yet uplifted by the Pliocene.

REGIONAL CROSS SECTIONS

The Transylvanian Basin is one of the major gas-producing areas of Romania (Paraschiv, 1979; Ionescu, 1994; Popescu, 1995). Due to extensive exploration and development of the post-salt Middle Miocene gas resource, a large subsurface database exists within the Romanian National Gas Company (ROMGAZ). It consists of more than several thousand km of 2D seismic profiles and more than 4000 wells. The relatively shallow gas fields (< 3000 m) produce biogenic gas, which originates in the post-salt succession. Only a few deep exploration wells have been drilled into pre-salt deposits and still fewer reached the basement (Table 1, Fig. 3).

The basin architecture was interpreted using regional seismic data and deep wells. E-W and N-S sections (Fig. 6, Plates 1-6), are presented as uninterpreted reflection profiles and as interpreted line drawings. The sections of the regional plates represent close-ups selected for this paper (Fig. 6).

UPPERMOST CRETACEOUS RIFT MEGASEQUENCE

Definition

The Uppermost Cretaceous synrift megasequence is bounded by the Pre-Santonian (post-Middle Cretaceous basement assembly) and pre-Paleocene (Late Maastrichtian inversion of rift systems) unconformities and/or their correlative conformity surfaces.

Seismic interpretation

Our regional seismic interpretation indicates three major Late Cretaceous depositional areas in the subsurface of the Transylvanian Basin as follows: Târnave, Puini and Alămor basins (Ciupagea et al., 1970; de Broucker et al., 1998) (Fig. 7).

The Târnave Basin is a Late Cretaceous half-graben system on the top of Transylvanides. The Transylvanides were thrusted during the Mid Cretaceous over the passive margin of Dacia (Fig. 2). Seismic profiles provide evidence for Late Cretaceous rifting of the Târnave Basin (Plates 3-6; Fig. 4, 8). The basin is bounded by major, gently-curved, overall NE - SW oriented normal faults (Fig. 7). Several small-scale normal faults have also been interpreted (Plates 3, 5; Fig. 4), as active during the early stages of Late Cretaceous extension. The extension clearly postdates the Early Cretaceous, as seen on seismic interpretation (Fig. 4, Plate 3). Two wells in the Târnave Basin penetrated deep-marine Upper Cretaceous deposits (Fig. 7).

The Puini Basin (de Broucker, 1998; Ciulavu and Bertotti, 1994) is a Santonian – Maastrichtian rift/sag basin, that seals the Aptian to Coniacian basement-involved Puini thrust sheet and its flexural basin fill (Fig. 9). Stratigraphic age control is represented by the Puini well (Fig. 9). Similar structures and deposits outcrop in the SE part of the Apuseni Mountains. (Bleahu et al., 1981; Schuller, 2004) (Plate 2). In that area, basement units (basaltic rocks conformably covered by Jurassic to Lower Cretaceous shallow-marine carbonates; Săsăran et al., 1999) are thrusted eastwards over deep-marine deposits (Aptian to Albian wildflysch and Cenomanian to Coniacian flysch). These deep-marine sediments were deposited in a flexural basin formed as a response of the coeval eastward-directed basement thrusting. Younger, Santonian to Maastrichtian sediments unconformably seal basement thrusts and conformably cover the flexural basin succession (Fig. 10).

North of the Puini Basin several minor half-grabens (Dej basins) were interpreted (Plate 6), as being controlled by roughly NS striking normal faults (Fig. 7). Similar structures outcrop in the Northern Apuseni Mountains (e.g. Roşia and Borod Basins; Bleahu et al., 1981).

Seismic interpretation also suggests deep erosion of crystalline nappes beyond the Late Cretaceous depositional area, suggesting major pre-Late Cretaceous uplift and exhumation. This is also well documented by fission-track studies (Bojar et al., 1998; Dallmeyer et al., 1999; Sanders, 1999; Willingshofer et al., 1999; Pană et al., 2002).

Sedimentary evolution

The Uppermost Cretaceous sediments of the Transylvanian Basin form a large-scale transgressive – regressive cycle involving deep to shallow deposits. The regressive part includes shallow-marine sandstones and continental red-beds of Maastrichtian age (Bleahu et al., 1981; Antonescu et al., 1983; Lupu and Lupu, 1983; Lupu et al., 1993; Codrea and Dica, 2005). These may be related to abrupt changes in basin subsidence triggered by Late Maastrichtian inversion (Schuller, 2004). Locally, the Uppermost Cretaceous is only in deep-marine facies. In this case, these sediments conformably cover earlier deep-marine successions (Bleahu et al., 1981; Lupu, 2002; Schuller, 2004) and are covered deep-marine Paleocene deposits (e.g. Puini Basin, Filipescu et al., 2005).

Uppermost Cretaceous deposits outcrop in the mountains around the Transylvanian Basin (Fig. 7). In contrast to the mostly shallow-marine Uppermost Cretaceous of the Transylvanian Basin that seals the Middle Cretaceous basement assembly, coeval deposits of the Eastern Carpathians foredeep consist of deep-marine turbidites and pelagic shales (Melinte and Jipa, 2005 and references therein) that are underlain by Jurassic to Lower Cretaceous turbidites presumably deposited on either oceanic or transitional crust (Fig. 10). Upper Cretaceous shallower-marine sediments have been deposited in wedge-top basins overlying the Carpathians accretionary wedge (Fig. 10).

Discussion

$^{40}Ar/^{39}Ar$ cooling ages of 117-100 Ma were reported from north-vergent ductile shear zones of the Apuseni Mountains basement nappes (Dallmeyer et al., 1999). The oldest cooling ages indicate that Mid Cretaceous nappe emplacement was accompanied by strong exhumation that clearly pre-dates the onset of Latest Cretaceous sedimentation (Dallmeyer et al., 1999; Sanders, 1999). Cooling ages at around 100 Ma associated with retrograde metamorphism (greenschist facies) and mylonitization were interpreted as being associated with large-scale dextral wrenching coeval with the Latest Cretaceous basin formation (Pană and Erdmer, 1994; Dallmeyer et al., 1999; Willingshofer et al., 1999).

The collision climaxed during the Mid Cretaceous (Săndulescu, 1988; Balintoni, 1994), but local shortening may have continued during the Cenomanian – Coniacian. Schuller (2004) interpreted this as an Early to Mid Cretaceous "hard" collision (uplift), followed by a Late Cretaceous "soft" collision, retreating subduction and rapid, locally focused subsidence. The final suturing probably occurred during the Paleogene (Mațenco et al., 2005) and was associated with small-scale basin inversion (De Broucker et al., 1998; Ciulavu et al., 2000). The Latest Cretaceous extension is probably related to the post-orogenic collapse of the over-thickened Middle Cretaceous orogen (Ciulavu and Bertotti, 1994; Sanders, 1999; Willingshofer et al., 1999; Ciulavu et al., 2000); thus this rifting was not related to conventional lithospheric stretching. The Latest Cretaceous extension in the shallower crustal levels was coeval with mild shortening on the deeper crustal levels (Dallmayer et al., 1999; Schuller, 2004).

The Santonian to Maastrichtian deposits of the Carpathians seal the Middle Cretaceous nappe contacts of the Internal Eastern Carpathians. They consist of deep-marine, reddish, fine-grained turbidites and suspension-fallout pelagic sediments (e.g. Melinte and Jipa, 2005) which overlie the Carpathian accretionary prism (wedge-top basins and foredeep basins). To the west (i.e. toward the Transylvanian Basin), they wedge out, becoming progressively shallower, as indicated by 1) the shelf deposits sealing the thick-skinned nappes of the Inner Eastern Carpathians (Fig. 10) and 2) by the wide erosional domain indicated by our seismic interpretation, yet further to the west (Fig. 8). This suggests that the thin-skinned accretionary prism of the Eastern Carpathians was submerged during the Late Cretaceous (Fig. 10).

PALEOGENE SAG MEGASEQUENCE

Definition

The Transylvanian Paleogene megasequence fills sag basins, and is bounded by the pre-Paleocene (Latest Maastrichtian) and pre-Lower Miocene unconformities and/or their correlative conformities toward the northern Transylvanian Basin. The Latest Maastrichtian unconformity is related to the inversion of the Latest Cretaceous rift systems. Final emplacement of the Pienides nappes, coeval with flexural basin formation, triggered the pre-Lower Miocene unconformity.

Seismic interpretation

Seismic data covering the basin document a Maastrichtian inversion of the Latest Cretaceous rifts, and possibly older structures (Plates 1-6; Fig. 8, 9, 11, 12). The Latest Maastrichtian unconformity may be traced as a correlative conformity toward the north Transylvanian basin and locally into the base of Paleocene sag basins (e.g. Puini Basin). The Latest Maastrichtian inversion and uplift introduced widespread continental environments during the Paleocene (alluvial fans, fluvial deposits; Jibou Fm; Fig. 5). (Fig. 11, Plates 1, 2).

The Paleogene succession includes two 3rd order sequence boundaries that appear to be related to minor inversions of the Paleogene basins (Proust and Hosu, 1996; De Broucker et al., 1998). These minor unconformities and their correlative basinward conformities divide the Paleogene megasequence into three separate evolutionary phases corresponding to Danian (?) – Early Priabonian (Sag 1), Late Priabonian – Early Rupelian (Sag 2) and Late Rupelian – Chattian (Sag 3).

On dip-oriented profiles the Danian – Early Priabonian (Sag1) shows a wedge-like set of downlapping reflectors overlain by a set of overall onlapping reflectors (Fig. 11, Plates 1, 2). This sequence consists of prograding Paleocene – Lower Eocene alluvial fans (LST), followed by Bartonian shallow-marine deposits (TST, HST) (Fig. 5, Fig. 11). The intra-Priabonian inversion (Plate 2; Fig. 12) triggered a relative sea-level fall, erosion and development of fluvial systems (LST).

The Late Priabonian – Early Rupelian (Sag2) is mainly a shallow-marine sequence organized as transgressive – regressive cycle (Fig. 5). Late Priabonian reflectors onlap the inverted Sag1 succession (Plate 2; Fig. 12). The Sag2 phase ends with the Early Rupelian inversion (Plate 2; Fig. 11, 12), as indicated by the subsequent Late Rupelian onlaps.

The Late Rupelian – Chattian (Sag3) locally onlaps the Early Rupelian unconformity (Plate 2; Fig. 11, 12). The onlapping deposits consist of lowstand and transgressive depositional settings with widespread anoxic environments (Ileanda Formation, Valea Carelor Formation). The Sag3 basin fill extensively covers various older units (Fig. 9, 11 Plates 1-6) underlining the strongly transgressive character of the Late Rupelian (De Broucker et al., 1998; Ciulavu et al., 2000). The Chattian is mainly siliciclastic and overall progradational (Fig. 5) with widespread continental environments. This change is coeval with a worldwide eustatic sea-level fall (Hardenbol et al., 1998) and climate cooling (Abreu et al., 1998).

Sedimentary evolution

In the hinterland of the Carpathians, Paleogene outcrops occur around the present day margins of the Transylvanian Depression (Fig. 6). Particularly in the northern parts of the Transylvanian Basin, widespread Paleogene deposits crop out in the Pienide nappes (Fig. 6). The Pienides are allochthons thrusted over the northern part of the Transylvanian Basin during the Early Miocene. They originate from a sedimentary basin situated to west of the Transylvanian Basin (Săndulescu, 1988; Csontos et al, 1992; Dicea et al., 1980a, 1980b; Györfi et al., 1999; Aroldi, 2001; Tischler et al., 2003). Strictly speaking, the Pienides are not part of the Transylvanian Basin, so they will not be addressed here.

The framework of the Paleogene stratigraphy and its cyclic sedimentation was established by the end of the 19th century (e.g. Koch, 1894). During the Paleocene – Early Eocene in the western and eastern part of the basin (Fig. 13), alluvial-fan and fluvial systems deposited coarse to fine-grained reddish sediments with several lacustrine intercalations (Gheerbrant et al., 1999; Bucur et al., 2001; Codrea and Hosu, 2001). Deeper marine environments have been interpreted in the Puini basin (Kaminski and Filipescu, 2005). In general, the sedimentary environments deepen toward the north (Fig. 13).

Most of the continental deposits were flooded by the Middle Eocene. The Middle to Late Eocene sedimentary succession consists of two transgressive-regressive cycles each composed of evaporites, shallow-marine carbonates, outer shelf marls, shallow-marine sands and, finally, fine-grained fluvial deposits (Călata and Turea Groups; Koch, 1894; Proust and Hosu, 1996).

By the Early Oligocene the carbonate-dominated Eocene sediments were replaced by siliciclastics. Most of the southern part of the basin was exposed (Fig. 14). Continental, inner shelf (Buzaş Formation) and outer shelf (Ileanda Formation) environments dominated the central part of the basin (Petrescu et al., 1989 and references therein; Rusu, 1995; Filipescu, 2001). Toward the north, deeper marine slope (Vima Formation), slope/outer fan (Valea Carelor Formation) and middle fan (Birțu Formation, Borşa Formation) settings prevailed (Mészáros et al., 1971; Dicea et al., 1980a).

The shallow-marine Oligocene forms a large-scale transgressive – regressive cycle (Fig. 5). This contrasts with the deep-marine Oligocene sedimentary trends, i.e. where the Oligocene is part of the large-scale Oligocene – Lower Miocene regressive succession. The differences in the evolutionary trends reflect different source areas: south to north for the shallow-marine and west to east for the deep-marine (i.e. the emerging Pienides; Contescu et al., 1966; Dicea et al., 1980a; Aroldi, 2001; Tischler et al., 2003).

Discussion

The Latest Maastrichtian inversion involved the reactivation of Late Cretaceous extensional faults (eg. Târnave Basin, Fig. 8) and older basement structures (e.g. the Puini thrust, Fig. 9, 11). The strong inversion is well illustrated by widespread Paleocene coarse-grained continental environments and high-energy depositional systems (alluvial fans). These contrast with the Eocene – Early Oligocene carbonate-dominated quiet sedimentary environments that were affected by only minor inversions and local unconformities. In the Paleocene – Eocene there is no evidence for any shortening or extension. Huismans et al. (1997) suggest a possible Early Rupelian N-S oriented extension. The Paleocene – Eocene subsidence may well be the result of crustal cooling following the mid-Cretaceous to Paleogene collision. Coeval N-S oriented extension and related simple-shear basin formation has been proposed by Săndulescu (1992, cf. Mațenco, 1997) for the Eastern Carpathian foredeep. In contrast, Zweigel et al. (1998) indicate only an overall NW-SE contraction and weak orogen parallel-extension. In the Southern Carpathians, Eocene WSW-ENE oriented orogen-parallel extension was documented (Linzer et al., 1998; Schmid et al., 1998; Mațenco and Schmid, 1999; Fügenschuh and Schmid, 2005).

Beginning with the Late Rupelian, major sedimentary and faunal changes occur in the Transylvanian Basin (Rusu, 1989; Popescu and Brotea, 1994). The tectonic regime suggests an ENE-WSW oriented compression (Huismans et al., 1997; Györfi et al., 1999). These changes indicate the onset of the Sag3 basin development.

The Oligocene regional compression may be related to the inception of the final closure of the Pienides domain (Györfi et al., 1999; Tischler et al., 2003). At least the northern part of the Transylvanian Basin was affected by weak flexural downbending as indicated by an overall northward thickening of the Oligocene succession and the western source of the Oligocene flysch in the northern Transylvanian Basin (Săndulescu and Micu, 1989; Aroldi, 2001; Tischler et al., 2003).

Other authors (e.g. De Broucker et al., 1998; Proust and Hosu, 1996; Ciulavu et al., 2000; Mațenco et al., 2005) interpreted the Paleogene Transylvanian Basin as a piggyback or foreland basin. We do not interpret the minor Paleogene inversions as the main trigger of Paleogene subsidence. The sedimentation pattern suggests a rather slow subsidence and quiet sedimentary environments. We also prefer to avoid the term "piggyback" because, following Ori and Friend's (1988) original definition, the Transylvanian Basin tectonic situation is quite different from the one described by these authors.

In contrast to the mainly shallow-marine/continental Paleogene of the Transylvanian Basin (Fig. 13, 14), the Eastern Carpathian Paleogene foredeep was entirely deep-marine (Fig. 5) (e.g. Săndulescu et al., 1987). The Paleocene – Lower Eocene consists of coarse-grained turbidites (ex. Tarcău sandstone), followed by a fine-grained Middle – Upper Eocene succession. The Priabonian consists of a coarsening-upward succession of deep-marine pelagic limestones and/or outer-fan shales, middle-, and proximal-fan sandy turbidites. The Oligocene comprises Lower Rupelian sandy turbidites, Upper Rupelian bituminous shales (locally called menilites) and Chattian coarse-grained sandstones deposited in deep-marine (Fusaru and partly Kliwa facies) and shelf settings (Kliwa) (Fig. 5, 15). The bituminous shales represent one of the most important hydrocarbon sources in the fold belt. Based on provenance studies, the Carpathian Paleogene foredeep had two major source areas: one in the east (the East European Plate) and other in the west (i.e. the Transylvanian Basin and/or Cretaceous Carpathian accretionary prism) (Săndulescu and Micu, 1989).

The Rupelian facies point to a relatively weak sedimentary input. This is coeval with deepening of the northern Transylvanian Basin, and both domains change to anoxic environments (bituminous shales) (Fig. 5). Seismic interpretation also suggests that the Oligocene thickens toward the eastern Transylvanian Basin and the southern parts were eroded. Therefore, at least the northern part of the Eastern Carpathian accretionary prism was partially submerged (Fig. 15), and we may visualize a low emerging ridge separating the southern Transylvanian Basin to the south from the main foredeep to the north. This concept is supported by the paleoflow directions of the Transylvanian Basin deep-marine Oligocene turbidites, which indicate an overall western to south-western source (Fig. 15), without any evidence indicating an eastern source (i.e. from the accretionary prism and its backstop).

Major sedimentary changes occurred during the Chattian, as indicated by the evolution of the Transylvanian Basin and the main Carpathian foredeep (Fig. 5). The foredeep was fed a by western as well as an eastern source (Săndulescu and Micu, 1989). Cretaceous to Eocene olistoliths that clearly originate from the accretionary prism are found in the Chattian flysch all along the Eastern Carpathians (Slon facies; Ştefănescu, 1980). These olistoliths suggest an actively rising accretionary wedge that separated most of the Late Oligocene Transylvanian Basin from the Carpathian foredeep.

LOWER MIOCENE FLEXURAL MEGASEQUENCE

Definition

The Lower Miocene megasequence is represented by the sedimentary fill of a flexural basin that developed in the central-northern parts of the Transylvanian Basin in response to the final thrusting and emplacement of the Pienides thrust nappes. This megasequence is bounded by the Lower Miocene basal unconformity, which can be traced toward the north (deeper part of the Transylvanian Basin) as a correlative conformity surface. The top of the megasequence consists of the pre-Middle Miocene basin-wide unconformity.

Seismic interpretation

On seismic sections, the Lower Miocene megasequence was mapped between the highest Oligocene and basal salt horizons (Plates 5, 6). For the upper boundary, we preferred to use the base salt instead of the top of the Lower Miocene, because the Lower Badenian (Dej Formation) is generally too thin to be resolved on seismic data and the Badenian/Lower Miocene boundary is ambiguous on seismic sections.

The Lower Miocene megasequence is wedge-like: it thickens toward the Pienides (thrust front) and thins toward the forebulge area (i.e. the central parts of the Transylvanian Basin), and includes distal, overall "downlapping" reflectors (Plates 5, 6). The "downlap" geometry is only apparent because we deal with an onlap surface that was tilted toward the SE during the Late Badenian to Pliocene (Krézsek and Filipescu, 2005). The depositional geometry can be readily reconstructed by flattening the base salt horizon (Fig. 16) and our seismic interpretation indicates a sedimentary architecture characteristic for asymmetric flexural basins (Fig. 17, 18).

At least three significant unconformities may be interpreted within the Lower Miocene megasequence. These unconformities delimit seismic packages with differing dips. The reflectors above the unconformities are amalgamated, have high-amplitudes and a channeled external shape suggesting coarse-grained sediments overlie the unconformity. The unconformities and related seismic facies changes indicate syn-sedimentary tectonic activity and the development of lowstand systems tracts.

Sedimentary evolution

The Lower Miocene outcrops mainly in the northern parts of the Transylvanian Basin (Fig. 6), and consists of deep-marine turbidites (Dicea et al., 1980a; Tischler et al., 2003) and coarse-grained fandeltas (Săsăran, 2003 pers. com)

grouped together in the Hida Formation (Koch, 1900) (Fig. 17). The upper part of the Hida Formation, excepting a few-fresh water molluscs (Popescu et al., 1995) and very scarce nannoplancton assemblages (Mészáros, 1991) lacks paleontological data with detailed chronostratigraphic value (Filipescu, 2001). Early Badenian deep-marine shales with rich planktonic fauna (Popescu, 1970) cover the Hida Formation. Well data (eg. 1 Strâmbu, 1 Coşbuc, 2 Telciu, etc.) and regional interpretations (Dicea et al., 1980a; Tischler et al., 2003) indicate that the Hida Formation is a large-scale shallowing-upward succession (Fig. 5).

In the western-central parts of the Transylvanian Basin, the Lower Miocene consists of transgressive littoral sandstones and outer shelf marls with rich faunas (Coruş and Chechiş Formation; Chira, 1994; Popescu et al., 1995; Filipescu, 2001). In these areas, the Lower Miocene appears to be truncated by pre-Middle Miocene erosion (Fig. 5), thus the uppermost part of the Lower Miocene is missing. The preserved succession consists mostly of fining-upward cycles that contrast the overall coarsening-upward trends of the north.

In the southern parts of the Transylvanian Basin, red continental and/or shallow-marine deposits have been locally reported (e.g. Ciupagea et al., 1970). There is an ongoing debate about the age of the continental deposits (Baluță, 1987; Grigorescu, 1987; Gheorghian and Gheorghian, 1994; Codrea and Dica, 2005). Their tectonic and sedimentary relation with the remaining Lower Miocene Transylvanian Basin also is not clear.

Discussion

The Lower Miocene Transylvanian Basin is a WSW to ENE striking asymmetric flexural basin tectonically loaded by southeast-ward thrusting of the Pienides at the eastern tip of the Alcapa terrane (Csontos, 1995; Györfi et al., 1999). The Alcapa terrane, like the Tisa-Dacia terrane, is yet another intra-Carpathian tectonic terrane assembled during the Mid Cretaceous that is located to the west of Tisa-Dacia (e.g. Csontos, 1995; Fodor et al., 1999). During the Lower Miocene the Transylvanian Basin was bounded to the north by the Pienides thrust system and to the south by an ill-defined forebulge in the central parts of the basin (Fig. 17, 18).

The unconformity and onlap surface at the base of the Lowest Miocene megasequence (e.g. Fig. 16) is due to the rapidly overriding tectonic load of the Alcapa – Pienides and the coeval initiation of a southward migrating flexural bulge. The combined effect of forebulge migration and increased subsidence led to transgression in the distal reaches of the flexural basin, as indicated by the deepening of the sedimentary environment from proximal (Coruş Formation) to outer shelf (Chechiş Formation). The fining-upward tendency of the Lower Miocene cycles is not as evident on northern well logs, as in the nearby source areas (i.e. Pienides thrusts) the sediment supply exceeded the accommodation space generated by the tectonic load. In the distal reaches of the flexural basin, shallow-marine (Coruş Formation, Chechiş Formation) and fluvial (Râpa Roşie Formation) sediments predominate. These contrast with the northern parts of the basin, where increased subsidence maintained earlier deep-marine environments, but, as compared to the Oligocene fan systems, the Lowest Miocene sediments are coarser-grained. The southern parts of the Transylvanian Basin were perhaps the backbulge region of the flexural basin, as suggested by thin local shallow-marine and continental deposits.

The upper part of the Lower Miocene megasequence is dominated by coarse-grained fan deltas. This suggests an overall decrease in subsidence rate and accommodation space due to the end of thrusting. Coeval isostatic relaxation of the crust leads to uplift and erosion in the forebulge area, as indicated by the missing upper part of the Lower Miocene in that area.

The Lower Miocene (Lower Burdigalian) Carpathian foredeep resembles the Oligocene sedimentary setting. The internal foredeep facies is represented by submarine-fan sandstones with slumps and slides that shale out toward the foreland (bituminous shales and outer fan turbidites). The thrusting of the Upper Cretaceous to Lowest Miocene deep-marine sediments of the foredeep was initiated during the Burdigalian. This is indicated by progressively shallower marine environments (evaporites and sandstones) on the top of the progressively growing Early Burdigalian fold belt (Fig. 17). The evaporites and shallow-marine sandstones are followed by Upper Burdigalian post-tectonic conglomerates and sandstones of wedge-top basins sealing the intra-Burdigalian nappe contacts.

The Carpathian accretionary wedge progressively emerged above sea-level at the beginning of the Paleogene, but was active mostly during Late Oligocene – Early Miocene, affecting mainly the northern parts of the Carpathians (north of the Trotuş fault; Maţenco, 1997; Tărăpoancă et al., 2003, 2004). This stage of active emergence corresponds to the main thrusting within the Paleogene to Lower Miocene foredeep.

To sum up, major changes occurred during the Early Miocene, both in the Transylvanian Basin and the Carpathians. However, the Transylvanian Basin megasequence development was mostly governed by the thrusting of the Pienides and only secondarily by the Carpathians.

MIDDLE TO UPPER MIOCENE BACKARC MEGASEQUENCE

Definition

The Middle to Upper Miocene megasequence is developed in the backarc region of the Carpathians and evolved within a regional compressional stress field. It is bounded by the pre-Middle Miocene and the post-Upper Miocene regional unconformities.

Seismic interpretation

One of the most characteristic seismic layers of the megasequence is formed by salt (Middle Badenian), which forms an almost continuous subsurface layer in the Transylvanian Basin (Ciupagea et al., 1970). The salt layer separates the Lower Badenian (pre-salt) from the Upper Badenian – Pannonian (post-salt) seismic packages (Plates 1-6).

Due to post-depositional salt-tectonics, the salt varies in thickness (Fig. 19) (Krézsek, 2004). The salt structures of the basin generally trend NW to SE, and include salt-cored anticlines (basin center) and two diapiric alignments near the eastern and western parts of the present Transylvanian Depression (Fig. 19, Plate 2) (Ciupagea et al., 1970). Salt tectonics will be discussed in detail in the following section.

The pre-salt seismic package is generally too thin (< 100 m; Ciupagea et al., 1970) for detailed seismic interpretation. Nevertheless, a very strong-amplitude seismic reflector is identified as the Dej Tuff horizon (Fig. 5), which represents one of the best widespread regional markers of the Transylvanian Basin (Plates 1-6).

The lowest post-salt succession (lower part of the Upper Badenian) onlaps the salt layer at the northern, western and southern parts of the basin and thickens toward the southeast (Plates 1-6, Fig. 20). Its thickness ranges between a few hundred meters in the west to more than 2000 meters in the east (Fig. 20). It is important to note that the present-day thickness distribution is closely related to the strong post-Middle Miocene uplift of the basin-bordering mountains that led to complete erosion of post-salt sediments from the basin margins (Sanders, 1999).

Several locally-developed post-Badenian unconformities have been recognized mainly in the central-eastern parts of the basin (Plates 3, 4) (Krézsek and Filipescu, 2005). Most of them are strongly related to salt tectonics.

The whole post-salt succession is gently folded (Plates 1-6, Fig. 24) suggesting post-Upper Miocene folding and thrusting. Reverse faulting is important in the western and southern parts (e.g. Ruși-Cenade fault) of the basin (Plates 1-5). This contrasts with the structural style east of the eastern diapir alignment (Plates 1, 2). Most of the post-salt faults terminate downward at the salt layer. This indicates that the post-salt seismic package is detached from the whole pre-salt succession, the salt representing a regional detachment layer.

Sedimentary evolution

The Middle Miocene (Lower Badenian) begins with a strongly transgressive event (Fig. 5) and an associated planktonic bloom (Popescu, 1970). In the western part of the basin, carbonates and marls (Gârbova Formation) were deposited in shallow to outer-ramp environments (Filipescu, 1996; Hosu and Filipescu, 1996; Filipescu and Gîrbacea, 1997). Deeper sedimentary settings (a few hundreds meters deep) in the central and eastern part of the basin are characterized by turbidites, pelagic microfossils and several rhyolitic tuff horizons (Dej Formation). The radiometric age of the Dej tuff is 15.6 My (Seghedi and Szakács, 1991). The uppermost Lower Badenian consists of deep-marine marls, which suggest deep-marine condensation (Filipescu, 2001).

During the Middle Badenian, shallow-water gypsum (Koch, 1900; Ciupagea et al., 1970) was deposited in the western (Gherghari et al., 1991) and southwestern parts of basin (Lubenescu, 1981), while in the remaining area deeper-marine salt formed (Fig. 5, 19) (Pauca, 1967; Dragoş, 1969; Ciupagea et al. 1970; Stoica and Gherasie, 1981). The evaporite distribution suggests a westwards shallowing sedimentary basin (Fig. 19, 21).

The post-salt succession is dominated by deep-marine clastic depositional systems (Fig. 5) (Krézsek and Filipescu, 2005). Shallow-marine sands and outer ramp marls are preserved mainly in the western and southwestern parts of the basin (Fig. 20) or they sporadically outcrop as part of lowstand systems tracts wedged between deep-marine sediments of highstand and transgressive systems tracts (Krézsek and Filipescu, 2005). The paleocurrent directions indicate that the submarine fans were fed by source areas located southwest and north to northeast of the Transylvanian Basin (Fig. 20). The sedimentation became overall regressive only in the Upper Miocene (Krézsek, 2005).

Most of the Upper Miocene basin fill has been eroded by the regional Pliocene uplift (Krézsek, 2005). Fission track studies (e.g. Sanders et al., 2002) indicate that at least 500 m to locally more than 1000 m of sediments have been eroded since the Pliocene in the Transylvanian Basin. Based on paleontological data the youngest preserved Upper Miocene deposits of the basin are around 9-10 My in age (Magyar et al., 1999; Magyar, 2005 pers. com) and are typically represented by deep lacustrine fans (Fig. 5) (Krézsek, 2005).

Intense magmatism during the Middle to Upper Miocene Carpathians deformation formed a volcanic arc behind the Eastern Carpathian accretionary wedge (Fig. 6). The volcanic activity started during the Sarmatian (~14 My) in the northern part of the Transylvanian Basin and progressively migrated in time and space towards the south, roughly following the eastern margin of the Transylvanian Basin (Pécskay et al., 1995; Szakács and Seghedi, 1995; Seghedi et al., 1998, 2004). The youngest volcanic rocks were described from the southern part of the volcanic chain and yield an age of 0.2 Ma (Pécskay et al., 1995; Szabó et al., 2004). The volcanic products cover most of the eastern margin of the Transylvanian Basin (Fig. 6) (Krézsek, 2005).

Discussion

The Middle to Upper Miocene megasequence began with slow regional subsidence and flooding of the Early Miocene continental environments that established shelf to neritic (a few hundred meters deep) sedimentary settings (Krézsek and Filipescu, 2005). This transgression is well known from other intra-Carpathian basins (see Royden, 1985; Vakarcs et al. 1994; Kováč et al. 1995; Tari and Horváth, 1995; Decker, 1996; Sachsenhofer, 1996). In contrast to those basins, the transgression in the Transylvanian Basin is not clearly related to extension (Krézsek and Filipescu, 2005; Tiliță

et al., 2006). Field data might indicate a weak earliest Badenian east-west oriented extension (Huismans et al., 1997), but none of this is evident from seismic data. We consider that, the earliest Badenian extension is not likely to be the main reason for the widespread subsidence because would be too weak to generate the observed subsidence pattern.

It is thought that the evaporites were deposited in a relatively deep basin formed by regional desiccation triggered by relative sea level fall (Krézsek and Filipescu, 2005). Coeval shallow evaporite facies consist of gypsum (Fig. 19, 21). It is important to note that just below and just above the salt only deep marine pelagic sediments have been recognized so far, without any clear evidence of shallow-water deposits (Popescu et al., 1995; Filipescu, 1996, 2001). In the Pannonian Basin, the ongoing backarc extension and related thermal uplift prevented salt deposition, but triggered widespread erosion (Báldi, 1980; Vakarcs et al., 1994; Tari and Horváth, 1995; Báldi et al., 2002) (Fig. 21).

Middle Badenian salt deposits are present in the Carpathian foredeep (Mrazec, 1907; Slaczka, 1987; Kasprzyk and Orti, 1998; Peryt, 2001; Ştefănescu et al., 2000). They cover Lower Badenian tuff horizons (Slănic tuff) that are coeval with the Dej tuff horizon of the Transylvanian Basin. Early paleogeographic reconstructions by Pauca (1967) suggested that the Transylvanian salt basin was widely connected to the Carpathian foredeep salt basin (Fig. 19) which stretched from Poland to Romania. It is worth mentioning that salt was also deposited during the Early Miocene. The Early Miocene salt in the Carpathian foredeep represents the most effective detachment layer of the Carpathian nappes (e.g. Ştefănescu et al., 2000 and references therein). The Middle Miocene salt was also involved in later deformations but represented a less-effective detachment.

The lower part of the post-salt succession (Upper Badenian) clearly onlaps the salt layer (Plates 1-6, Fig. 20). In contrast to previous interpretations (e.g. Krézsek and Filipescu, 2005), newly acquired data suggest that the onlapping sediments are most probably shelf-type deposits that change basinward into turbidites and bituminous shales (Krézsek et al., unpubl. data). The deep-marine planktonic assemblages (Popescu et al., 1995; Filipescu, 1996) suggest no more than 1000 water depth. The bituminous shales lack benthonic assemblages and indicate anoxic conditions in the deepest marine settings. They are the primary source of the Transylvanian Basin biogenic gas accumulations. A change of microfaunas in the upper part of the Late Badenian (Filipescu, 2004) and extensive deep marine settings indicate wide connections with the Carpathian foreland and Pannonian Basin (Báldi et al., 2002; Tărăpoancă et al., 2003, 2004; Krézsek and Filipescu, 2005) (Fig. 22).

When compared with the Middle Badenian salt, the Upper Badenian deep-marine settings suggest a very strong post-salt deepening related to increased subsidence rates. The subsidence rates were not uniform across the whole basin, as suggested by the Upper Badenian thickness distribution, i.e. the thickest (up to 1500 m) and coarsest deposits are located in the southeastern part of the basin, while in other parts only a thin (100 – 500 m) and fine-grained succession is known (Fig. 22) (Krézsek and Filipescu, 2005). The Upper Badenian facies distribution may be interpreted as the effect of eastward increasing subsidence rates triggered by post-salt southeastward tilt of the basin (Plates 5, 6) (De Broucker et al. 1998). The subsidence rates remained high throughout the Middle Miocene (Crânganu and Deming, 1996).

The Middle to Upper Miocene megasequence documents a two-stage subsidence history: Early – Mid Badenian (Langhian) slow subsidence was followed by Late Badenian to Pannonian (Serravalian to Early Tortonian) rapid subsidence (Crânganu and Demming, 1996; Krézsek and Filipescu, 2005). The high subsidence rates and coeval tilting of the basin may have been related to the roll-back of the subducted slab which pulled down the basin basement (e.g. Royden, 1988, 1993; Cloetingh et al., 2004).

During the Late Middle Miocene, the Tisa-Dacia terrane collided with the East European foreland. This final collision uplifted the Carpathian Mountains that border the basin (e.g. Săndulescu, 1988; Maţenco, 1997; Gîrbacea and Frisch, 1998; Zweigel et al., 1998; Chalot-Prat and Gîrbacea, 2000; Maţenco and Bertotti, 2000; Tărăpoancă et al., 2004; Cloetingh et al., 2004) and lead to a rejuvenation of source areas and an increase in sedimentary input (Krézsek, 2005). This coupled with the overall decrease of the subsidence-generated accommodation space, triggered overall regressive sedimentation in the Late Miocene Transylvanian Basin (Fig. 5).

The roll-back of the slab associated with the above-mentioned collision led to breaking off the subducted slab, which diminished the pulling force and finally triggered isostatic uplift and related regional erosion of the basin (e.g. Cloetingh et al., 2004). The uplift age is Latest Miocene – Early Pliocene. The age of the erosion is constrained by the youngest preserved Upper Miocene deposits (~9 My) and by Upper Pliocene volcanics (< 3 My) that seal the erosional unconformities in the SE part of the basin. The Pliocene uplift was coeval with subsidiary extensional basins (Gîrbacea and Frisch, 1998; Chalot-Prat and Gîrbacea, 2000) opening the southeastern Transylvanian Basin, across the crest of the Carpathians (Fig. 23).

MIO-PLIOCENE GRAVITATIONAL SALT TECTONICS

The origin of the salt structures in the Transylvanian Basin has been of great historical interest and their major characteristics have been addressed by several authors (Pošepný, 1871; Koch, 1894; Mrazec, 1907; Pauca, 1967; Visarion et al., 1976; Stoica and Gherasie, 1981). However, only a few attempts have been made to interpret the evolution of the regional salt tectonics (e.g. Krézsek, 2004; Szakács and Krézsek, in prep.). In the following, we will show that the Transylvanian salt structures differ significantly from the Carpathian salt structures and superficially resemble the gravity-driven structures on passive margins.

In order to illustrate the overall setting of the salt structures of the Transylvanian Basin, regional cross sections (Plates 1-6) were constructed using seismic lines. Seismic lines on Plate 2 and Fig. 27 were extended further to the east using outcrop and well data, to show the relation between the Transylvanian Basin basement and the outcropping Eastern Carpathian basement.

The main salt-tectonic features of the basin are as follows:

1. In the far eastern part of the regional sections (Plates 1, 2; Fig. 26, 27) a pronounced extensional weld is associated with a west-dipping growth fault. Post-salt, internal detachments are also observed. The basement slope increases markedly toward the east. This slope can be extended further up dip by well control (1LuncaB, 1Zebrac; Plate 2), and appears to link up straight with the basement outcrops to the east (Fig. 24). In the eastern Transylvanian Basin, Bucur et al. (1971) mapped and illustrated in detail the outcropping Middle to Upper Miocene deposits (Fig. 25). They dip northeastward, at high angles (from 45 to 10 degrees).

2. In the eastern part of the sections, diapirs, which are more than 3 km high are observable. They form a NW to SW striking and NE dipping belt (Eastern Diapir Belt). Most of the flanks of passive diapirs are cut off by reverse faults induced by the late shortening (Plate 1, 2). However, the discordant contacts are not merely fault cut offs but suggest earlier reactive/passive diapir growth (e.g. Vendeville and Jackson, 1992).

3. In the center of the sections, a number of salt-cored folds are visible (i.e. central fold domain). The post-salt succession anticlines represent salt-cored folds or thrusts, and some of them are faulted folds, which dip slightly to the east. The undulating anticlinal axial traces of most of the anticlines and synclines (Fig. 24) have been taken from the geologic map of Romania and supplemented by some structures that are shown on seismic sections. They roughly follow a very characteristic north-south trend, as is also suggested by the subsurface distribution of salt structures (Fig. 19, 24). The trend of the structures indicates a regional Mio-Pliocene east to west oriented maximum compressive stress, however this regional compression is not well defined by field data. Alternatively, we propose that they may be the result of large gravitational spreading on the eastern flank of the basin and off the slope of the rising Carpathians to the east. This concept is supported by the absence of significant folding in the pre-salt strata, which suggests that only the post-salt succession was involved in the deformation. The salt represented the regional décollement of the gravity spreading.

Sedimentary thinning above the salt-cored folds suggests slow rise (weak shortening) during the Mid Miocene (Late Badenian-Early Sarmatian) that generated salt pillows and gentle folds. These early folds were highly amplified by the post-Late Miocene shortening.

4. In the western part of the basin, salt extrusion is related to the Mio-Pliocene thrust system (Plates 1, 2, Fig. 12). Seismic interpretation does not support the presence of pre-shortening diapirs or other salt tectonics-related weak zones in the far west (e.g. the presence of syn-kinematic thickening/thinning). The thrust systems represent the toe of basin-scale gravitational slide as it abuts against the subsurface flank of the Apuseni Mountains but may also be related to the Pliocene uplift of the Apuseni Mountains. Seismic evidence in the subsurface of the Transylvanian Basin for some very late basement-involved thrusting associated with that broad uplift is intriguing but not compelling. Seismic interpretation suggests that the detachment is folded together with underlying basement (e.g. Plate 3), thus the toe of spreading is tilted by the Apuseni uplift. Concentration of shortening in the far west can be the result of salt to gypsum facies change that occur near the Apuseni Mountains, thus tapering of the ductile layer. In this case, the central overburden of the basin was weakly folded while passively translated on the salt.

5. In the southern part of the basin, an important east to west striking fault is known as the Ruşi-Cenade fault (Fig 24; Ciupagea et al., 1970; Ciulavu et al., 2000). The northward dipping fault thrusts Sarmatian strata on top of Pannonian (Fig. 28). Tectonic transport directions indicate thrusting to SW (Ciulavu et al., 2000). Seismic data show that the reverse faulting is not related to basement-involved thrusting and affects only the post-salt succession (Fig. 28). We interpret this to be the compressional southern lateral ramp of the slide system. This is supported by the different architecture of the salt-structures to the north compared to the south of the thrust fault, i.e. the north-south striking regular pattern of the salt-cored folds stops south of the thrust fault (Fig. 24). Initiation of Ruşi-Cenade fault was controlled by salt distribution that was thinner toward the south of the fault, and by a basement step that enhanced initiation of diapirism, thus facilitated thrusting (e.g. Ge et al., 1997).

6. In the northern parts of the basin, the northern lateral ramp of the slide is not preserved, due to greater uplift and erosion. However, we assume that it was linked to the northernmost salt outcrops of the basin (Fig. 24).

To the east the Pliocene volcanic deposits follow the overall eastward-dipping trend. The wide volcanoclastic plateau, developed preferentially westward by adjoining composite volcanoes, extends deep into the Transylvanian Basin and its base is tilted toward the chain axis (Fig. 25, 26, Plate 1, 2) (Krézsek, 2004; Szakács and Krézsek, in prep.). Upper Miocene outcrops in the central part of the volcanic range (e.g. Szakács and Seghedi, 1996), indicate a very shallow pre-volcanic basement, close to the present erosion base. Well 1Zebrac drilled in that area, traversed a flat lying Middle to Late Miocene sedimentary succession and have found no volcanics, nor any salt. This is in contrast to the eastward tilted post-salt succession of the 1LuncaB well (Plate 2). These data suggest that, eastward of well 1Zebrac (Fig. 24) the volcanic edifices and the post-salt overburden are barely tilted. Therefore the main detachment scarp of the gravitational spreading system is constrained between the 1LuncaB and 1Zebrac wells (Plate 2; Fig. 24). This is supported by many west-dipping extensional faults measured in outcrops between Lunca Bradului and Zebrac that contrasts with the area east of Zebrac, where no faults have been found (Fieltz and Seghedi, 2005). These faults offset volcanics younger than 7-8

Ma. The fault scarp outcrops farther north, as indicated by seismic and outcrop data (Fig. 24, 27). Further south, the scarp is inferred from seismic sections and inclined dips towards the east that can be followed all the way to the edge of the volcanics.

Discussion

Seismic, outcrop and borehole data suggest Mio-Pliocene gravitational spreading of the post-salt succession. The salt layer represents the décollement layer of the gravity spreading that is pretty well confined to the area underlain by salt. This gravitational slide occupies much of the central Transylvanian Basin (Fig. 24).

We interpret this as a gravitationally driven linked compressional – extensional system that was continuously active during the Late Mid Miocene – Pliocene (or even younger?). The post-Late Miocene gravity spreading was stronger than that during the Mid Miocene, as shown by the salt-cored folds of the central Transylvanian Basin.

In order to trigger the gravitational spreading, the surface of the overburden must dip sufficiently for the gravitational body force to be resolved into a downslope shear stress that is large enough to overcome the shear strength of the décollement layer or the frictional strength of the décollement fault (Jackson and Talbot, 1991). This topographic slope was progressively created as the Carpathians and the Transylvanian Basin eastern margin began to rise. Fission-track data indicate that the main uplift of the Carpathians started in the Late Mid Miocene (Sanders, 1999), and since then at least 5000 m of sediments have been eroded. The uplift rejuvenated sediment source areas indicated by the rapid progradation of coarse-grained fan delta and submarine fan systems (Krézsek and Filipescu, 2005). The rapid sedimentation enhanced salt tectonics by differential loading.

We conclude that the postulated Carpathian accretionary wedge uplift is coeval with the salt-based gravitational spreading. This observation suggests that we are not dealing with a Late Miocene flexural depression in the east side of the Transylvanian Basin (Sanders, 1999; Sanders et al., 2002). The gravitational spreading welded sediments originally overlying the salt to its base in the east of the basin (extensional salt domain). The salt weld is located beneath the western part of the volcanic edifices (Fig. 24). The reduced salt was expelled toward the west and squeezed to the surface westwards of volcanic edifices, creating salt walls (Fig. 19, 24). The pre-shortening distribution of extensional salt diapirs controlled the position of the salt walls. The pattern of the salt-cored fold lineaments (Fig. 24) found in the central Transylvanian Basin is related to the compressional stress created by the mostly late gravitational spreading. In the western part of the basin the observed thrusting and related salt extrusion may represent the toe of basin-scale gravitational slide as it abuts against the subsurface flank of the Apuseni Mountains. These thrusts appear to be closely related to the primary wedge out of the salt. However, the strong regional eastward tilting and erosion of the Miocene section of the western/northwestern Transylvanian Basin is likely to be related to an overall uplift of the Apuseni Mountains, which was roughly coeval with the uplift of the Eastern Carpathians.

Overall, the evidence supporting our interpretation should speak for itself. However, there is more work to be done to fully evaluate the scope of early salt deformation that may perhaps be related to reactive diapirism (as suggested to us by M. Jackson). Such diapirism may or may not be associated with an early inception of gravitational gliding. Furthermore, there are also good reasons for assuming that an early Apuseni high controlled the western edge of salt deposition and that later stages of the Apuseni uplift further influenced the western boundary of the gravitational spreading. In this context there is some evidence that this late uplift may have been associated with some basement-involved faulting.

During the Pliocene, the ongoing basement uplift was coeval with active backarc type calc-alkaline volcanism (Pécskay et al., 1995). The volcanics covered the eastern part of the basin (Fig. 24). Their weight considerably enhanced the post-salt overburden load and thus contributed to the late stage spreading. However, volcanic thickening of the cover would have strengthened and resisted deformation. The high heat flow generated by the volcanic activity enhanced the gravitational spreading by increasing the plasticity of the salt, and decreasing the shear strength of the décollement layer. The most important late spreading phase postdates the volcanic activity because the volcanoes are tilted together with the post-salt sedimentary cover. The gliding led to volcano-spreading processes and highly influenced the morphology of the volcanoes (Szakács and Krézsek, in prep.).

The salt tectonics of the Transylvanian Basin differs from the classic salt tectonics of Mrazec (1907), as his diapirs occur in the compressional context of the Southern Carpathians foreland thrust - i.e. accretionary wedge-belts (Paraschiv, 1979; Ştefănescu et al., 2000). he Transylvanian Basin salt tectonics shares with more conventional salt-based foreland folded belts the fact that folding is pretty well confined to the area underlain by salt (e.g. Harrison, 1995; Costa and Vendeville, 2002). The relative irregular pattern of the structural trends of Transylvanian or the Ebro Basins (Sans and Vergés, 1995) suggests lesser shortening in contrast with the regular pattern of folds, thus greater shortening of Melville Island (Harrison, 1995) or Zagros Fold Belt (Talbot and Alavi, 1996). The architecture of the Transylvanian Basin salt structures resemble the upslope extensional and downslope compressional domains of gravity driven fold belts developed on passive margins (e.g. Wu et al., 1990; Peel et al., 1996; Tari et al., 2002; Fort et al., 2004). However, we emphasize that the triggering mechanisms are quite different.

PLIOCENE TO QUATERNARY BASEMENT-INVOLVED THRUSTING

The latest deformation of the Transylvanian Basin is associated with the east flank of the Apuseni Mountains, where we observe that the toe of the Pliocene décollement system is uplifted together with the underlying basement (Plate

2, 3). Further to the north, a major N-S striking basement-involved reverse fault (Meseş thrust, Fig. 6) and several compressional folds are evident. The faulted Pannonian strata on the western hanging wall block of the fault show the post-Late Miocene activity of the fault. Therefore, we speculate that this basement-involved compressional system is the result of late-stage (Pliocene or Quaternary?) inversion and uplift of the Apuseni Mountains (Maţenco and Cloetingh, 2006). This inversion reflect compressional intra-plate stresses in the intra-Carpathian area due to the ongoing movement of the Adriatic indenter (Bada, 1999; Maţenco and Cloetingh, 2006).

This late (Pliocene to Quaternary) compressional system is at least coeval with the late spreading of the post-salt succession to the east. Therefore, during the latest development of the back-arc Transylvanian Basin, we recognize two major, at least partially decoupled, deformation systems:

(1) Gravitationally linked shallow extension and compression related to the westward spreading of the post-salt succession away from the uplifting Carpathians.

(2) Basement-involved compression in the western part of the basin and the Apuseni Mountains related to the inversion of the intra-Carpathian area.

RELATIONSHIP OF MIDDLE TO UPPER MIOCENE TRANSYLVANIAN BASIN TO FORELAND FOLD BELT

It is important to relate Transylvanian Basin tectonics to the Carpathians and the Apuseni Mountains (e.g. Horváth and Royden, 1981; Burchfiel and Royden, 1982; Royden, 1993). The setting of the pre-Middle to Upper Miocene megasequences has been discussed earlier, so this section will limit itself to post Early Miocene regional tectonics. To provide further insight, some greatly simplified balanced cross sections were constructed (Plate 7) that are based entirely on the excellent set of cross sections published by Ştefănescu et al. (1988) and Maţenco (1997).

Balanced cross sections

In this first approximation, none of the many available commercial balancing programs were used. Thus, in principle, the cross sections could be refined using modern computer technology and more advanced modern methodology. In the following, we present some of the reasons that suggest that, because of the many data-based uncertainties inherent in the method, a more rigorous approach may not be warranted. The aim here is to estimate an overall magnitude of shortening and to relate the kinematics of the foreland folded belt to the post-Lower Miocene subsidence and the salt tectonics of the Transylvanian Basin.

There are good reasons to accept the excellent mapping done by generations of outstanding field geologists in Romania and to also accept the detailed kinematics of the foreland folded belt (Săndulescu, 1988; Zweigel et al., 1998; Maţenco and Bertotti, 2000), which is clearly displayed on the maps and the regional cross sections of Romania (Ştefanescu et al., 1988). Thus, the regional sections presented here merely suggest different interpretations at depth. The reason for this is due to the paucity of data supporting the existence of one or more retro-thrusts where postulated by Sanders (1999). Our search is for one or more other explanations of the westward-dipping “basement” slope that is suggested by the seismic data, a few critical wells and the basement outcrops to east of our seismic sections. Unfortunately, much critical information is now buried under Late Neogene volcanics, which obscure the transition between the Eastern Transylvanian Basin and the west flank of the Eastern Carpathians.

Modern balanced cross sections were introduced in the sixties in Canada (Bally et al., 1966; Dahlstrom, 1969). Subsequent publications greatly refined the methodology, proposed by Dahlstrom (e.g. Perry et al., 1984). The reason for the innovation was that in the Alberta Foothills and Rocky Mountains explorers established for the first time that a gently mountain-ward dipping basement could actually be mapped with considerable confidence using reflection seismic profiles (and wide-angle refraction techniques or broadside refraction). Similar evidence became later available in many foreland folded belts of the world. In all cases, the dip and the actual depth of the top basement was quite sensitive to seismic velocity assumptions that were used for depth conversions i.e. using faster velocities in the inner zones of folded belts would steepen the basement, lower velocities would make the basement dip gentler. It turns out that for “balanced cross sections” the actual basement slope is one of the most sensitive parameters.

In some folded belts, such as the Carpathians or the Apennines, a reliable basement top cannot be readily defined seismically and/or geologically. First, the nature and age of the basement itself is unknown and secondly there is no adequate observable impedance contrast between the basement and the overlying sediments. In addition, in the Eastern Carpathians, a complicated structural evolution of the foreland “basement“ can be anticipated just from trying to connect the trends of the Dobrogea area with the structural trends of the Polish trough (e.g. Ziegler and Horváth, 1996). Lastly, it is known from the Carpathian foreland, as from many other foreland folded belts, that very significant mountainward-dipping normal faults offset the underlying basement. These faults are syn-orogenic but affect mostly the foreland. An intuitive explanation for these faults being associated with lithospheric flexure is often suggested. Obviously, when these faults become very prominent, it is no longer acceptable to extrapolate an undisturbed basement top under the adjacent folded belt. Some authors (e.g. Bally et al., 1986 in Italy) attempted to avoid all the previous problems by assuming a “putative basement top”, which served as an aid for balanced cross sections. This is convenient but obviously questionable and such a procedure always ought to clearly flagged as an uncertainty, hopefully with an explanation of the

circumstances. The slope of such "putative basement tops" would be determined by a reasonable extrapolation of the foreland basement top, perhaps corrected for and flattened somewhat allowing for the effect of known late to post-orogenic uplift of the whole mountain range. Such uplifts were known for a long time from many mountain ranges but today they are much better constrained by apatite fission track studies, and a host of radiometric dating methods (e.g. Sanders 1999). In general, given a known thickness for the sediments involved in the folded belt it can be said that the steeper the assumed basement slope, the larger the total amount of displacement of the overlying section. Conversely: lesser the slope, the smaller displacement. In the Romanian Carpathians the nature and the dip of the basement is shown on many cross sections (Ştefănescu et al., 1988; Maţenco, 1997), but unfortunately documentation based on reasonable quality seismic cross sections or else by deep wells is generally not available.

Yet another significant uncertainty is the general absence of detailed stratigraphic measured sections that would provide thickness information for all formations but particularly for the deepest sections involved in the deformation. In the absence of such information, there is no choice but to use the thicknesses used for the cross sections constructed by local experts (e.g. Bleahu in Burchfiel, 1976; Paraschiv, 1979; Săndulescu, 1988). However, the uncertainty increases greatly if the nature and thickness of the deeply buried stratigraphy remains essentially unknown. Thus, in Eastern Carpathians, the existence and thickness of a deep Jurassic section or the involvement of older Triassic sediments is essentially unknown. The end effect of this is that with a given basement slope the thicker the assumed deep stratigraphic section is, the less displacement will be evident. On the other hand, if the assumed deep section is thin, the overall displacement will increase considerably.

A final uncertainty is associated with the selection of the direction of the line of cross section, which ideally ought to coincide with the path of tectonic transport. For the southeastern Carpathians Zweigel et al. (1998) and Maţenco and Bertotti (2000) have proposed some directions of transport that coincide reasonably well with the cross section directions used by Ştefănescu et al. (1988). But here again there are significant uncertainties as the inner units may be more rotated than the outer units of a foreland folded belt. Units may also be rotated in opposite senses etc. Of course, in Romania any rotation of the crystalline "backstop" (i.e the old "traineau écraseur", the modern "indenter") would further add to uncertainties (e.g. Panaiotu, 1998; Dupont-Nivet et al., 2005 and references therein).

In view of the many uncertainties listed and many more, it may be reasonably asked whether balanced cross sections are sufficiently instructive to justify the time-consuming effort involved in constructing them. Perhaps in our case the main incentive for doing such reconstructions is:

a. to get a sense of the shortening magnitudes of a foreland folded belt, which overall had remarkably little impact on the Neogene deformation within the Transylvanian Basin. This relationship suggests that coeval gravitational deformation in the basin was essentially decoupled by salt from much of the foreland folded belt, while riding on the top of crystalline units of the Inner Carpathians and their partially inverted Upper Cretaceous and Paleogene basins.
b. to explore possible alternatives to Sanders's "retro"-reverse fault system that may supplement his fission track uplift ages, while at the same time proving the uplift responsible for the salt-based Upper Miocene "megaslide" that affected much of the center of the Transylvanian Basin.

It could easily be shown that alternative equally well "balanced" cross sections respecting all available data could be constructed to explain both Sanders's (1999) uplift as well as the Neogene Transylvanian megaslide. A more definitive solution will not become available unless modern high-quality industry reflection seismic profiles will be acquired for key areas in the Romanian foreland folded belt. The schematic sections shown on Plate 7 here, suggest that modern reflection seismic surveys may eventually lead to very substantial, but deep, new gas discoveries in this foreland fold belt.

Sedimentary evolution of the Carpathians foredeep

Similar to the Transylvanian successions the Middle Miocene of the East Carpathian fold belt is characterized by Lower Badenian transgressive shales and tuff horizons, Middle Badenian evaporites, Upper Badenian deep-marine shales and turbidites. The Lower Sarmatian is relatively fine grained (e.g. Tărăpoancă et al., 2003).

Based on facies distribution and paleoflow indicators of the Carpathian foredeep and the Transylvanian Basin (Tărăpoancă et al., 2003; Krézsek and Filipescu, 2005), we infer that the Early Badenian and more important, the Latest Badenian relative sea-level rises submerged large areas of the Lower Miocene Carpathian accretionary wedge. Therefore, the Carpathians did not represent an important sedimentary source before Late Sarmatian either for the foreland or for the back-arc area (Răbăgia and Maţenco, 1999; Maţenco et al., 2003; Tărăpoancă et al., 2003; Krézsek and Filipescu, 2005). During the Middle Miocene (Upper Badenian – Middle Sarmatian) at least, the present-day Carpathian bend was submerged (Fig. 22), while the northern part of the Eastern Carpathians and the western part of the Southern Carpathians may have formed a rather low elevated ridge (Fig. 20).

The Transylvanian Basin uplift began by Late Sarmatian (~12 My) and led to rapid Upper Miocene (pre 9 My) filling of the basin (Krézsek, 2005). This is in contrast to the evolution of the Carpathians foreland, where subsidence was still active during the Quaternary (Bertotti et al., 2003; Tărăpoancă et al., 2003, 2004; Cloetingh et al., 2004). The

Transylvanian Basin had been already uplifted by 9 My and eroded during marine sedimentation in the Carpathians foredeep (Fig. 23)

Discussion

Two balanced cross sections (Plate 7) were constructed. When comparing the two sections, the different amount of shortening is readily observable. Our reconstructions indicate that the overall shortening in the Outer Eastern Carpathians ranges between 50 km (Section 2) to at least 250 km (Section 1). The estimated amounts are comparable with those reported by others (Ellouz and Roca, 1994; Roure and Sassi, 1995; Maţenco, 1997). Generally, the Late Sarmatian in-sequence thrusting is limited to around 10 km shortening. Bearing in mind all the uncertainties listed above, in our reconstructions the larger amount of shortening (Section 1) reflects a deeper "putative basement" top.

Sedimentary evolution and fission-track uplift ages (Sanders, 1999) indicate that the Eastern Carpathians began to rise post-Middle Sarmatian. Since then, at least 2-5 km of sediments have been eroded from the rising Carpathians. This amount is comparable to the foreland molasse thickness (Plate 7). Based on work of Sanders (1999) in the area of Section 2 the uplift ages yield around 12 Ma and in the area of Section 1 only 4 Ma (Fig. 24).

Important differences exist also between the post-Middle Sarmatian molasse thickness distribution (e.g. Tărăpoancă et al., 2003). North of the Trotuş fault (e.g. Section 2) the molasse is at most 4 km thick, while south of the Trotuş fault it exceeds 8 km. The Trotuş fault represents an important crustal-scale fault separating two "underthrusted" crustal blocks (East European Block and Moesian Block) (Fig. 6). The different rheological properties of the underthrusted blocks may have had major influence on the architecture of the mountain uplift and foredeep development (Cloetingh et al., 2004). Collision with the cold and thick lithosphere of East European Plate led to a limited amount of shortening in central and northern Eastern Carpathians, uplift starting around 12 Ma, erosion of up to 5 km of sediments and the development of a 3-6 km foredeep. This contrasts with the subduction of the thin and re-heated lithosphere of the Moesian Block (south of Trotuş fault), characterized by decoupled mantle and crustal deformations, uplift by 4-0 Ma, post-collisional erosion of 2-4 km of sediments and development of a 9 km unusual deep foredeep basin (Cloetingh et al., 2004; Tărăpoancă et al., 2004).

The postulated uplift ages of the Eastern Carpathians indicate that the central-northern parts of the Eastern Carpathians were uplifted earlier then the southern parts. Therefore, the basement slope of the northern and central eastern Transylvanian Basin was much steeper then the basement slope the southeastern Eastern Carpathians. This fits well with the NNW to SSE orientation of the major salt structures of the Transylvanian Basin, which requires a basement slope dipping to the WSW (Fig. 24).

The Carpathians uplift was coeval with obvious along-arc migration of backarc volcanism from NW to SE (Pécskay et al., 1995), accompanied by decreasing height of the volcanoes, volume and complexity of the volcanics in the same direction (Fig. 24) (Szakács and Seghedi, 1995). Therefore, in the central parts of the volcanic arc, the volcanic load and heat flux was higher then in its southern part, thus enhancing gravitational spreading (Szakács and Krézsek, in prep.).

SUMMARY AND CONCLUSIONS

The Transylvanian Basin represents a post-Cenomanian sedimentary basin, locally comprising more than 5 km sediments, developed in the hinterland of the Carpathian subduction zone on top of thick-skinned Middle Cretaceous basement nappes. The basin evolved during post-Mid Cretaceous thin-skinned shortening in the Carpathian thrust and fold belt.

The sedimentary fill may be divided into four tectonostratigraphic megasequences: Upper Cretaceous (rift), Paleogene (sag), Lower Miocene (flexural) and Middle to Upper Miocene (backarc dominated by gravitational tectonics).

Late Cretaceous rift sediments unconformably seal the Mid Cretaceous nappe contacts and locally the syntectonic sediments deposited in small flexural basins. Rift subsidence was governed by the extensional collapse of the over thickened Middle Cretaceous orogen (e.g. Willingshofer et al., 1999). Uppermost Cretaceous sediments of the Transylvanian Basin form a large-scale transgressive – regressive cycle composed of shallow- and deep-water deposits. They contrast with the coeval, deep-marine deposits of the Carpathians foredeep. Paleogeographical reconstruction suggests that the thin-skinned Carpathian accretionary prism was submerged during the Late Cretaceous.

The Late Cretaceous rifts were inverted before the Paleogene. Post-inversion Paleogene sediments were deposited in sag basins. Based on two minor intra-Paleogene inversions the Paleogene succession may be further divided into 3 sequences (e.g. Proust and Hosu, 1996), as follows: Danian (?) – Early Priabonian (Sag 1), Late Priabonian – Early Rupelian (Sag 2) and Late Rupelian – Chattian (Sag 3). In contrast to the shallow-marine/continental Paleogene deposits of the Transylvanian Basin, the Eastern Carpathians Paleogene flexural foredeep basin was entirely deep-marine. Sedimentary facies and paleoflow directions indicate that during the Eocene? – Oligocene at least the northern part of the Eastern Carpathians accretionary prism was submerged. The Carpathians underwent an important uplift phase by Late Oligocene-Earliest Miocene time.

The Lower Miocene megasequence represents the sedimentary fill of a flexural basin developed in the central-northern part of the Transylvanian Basin as a response to emplacement of the Pienides thrust nappes. The Carpathians did

not control deposition of this sequence. The Early Miocene Carpathian foredeep basin resembles the Oligocene deep marine sedimentary setting. Changes occurred only during the late Early Miocene to shallow marine settings on growing Carpathian fold belt system.

The Mid-Miocene continental collision in the Eastern Carpathians led to uplift of the basin margins together with the rising Carpathians (Sanders, 1999). This coupled with the increased differential load and high-heat flow induced by Late Miocene to Pliocene backarc volcanism triggered large-scale Mio-Pliocene gravitational spreading of the post-salt succession (Krézsek, 2004; Szakács and Krézsek, in prep.). This megaslide comprises three structural domains, as follows: extensional weld (upslope), contractional folds (central) and contractional toe thrust (downslope).

In the Pliocene and coeval with the erosion of the Transylvanian Basin, small extensional basins opened on top of the Carpathians. Slab detachment triggering localized extension and subsidence was proposed to explain the evolution of these basins (Gîrbacea and Frisch, 1998).

The latest deformation of the Transylvanian Basin (western part) is associated with the Late Pliocene (Quaternary?) basement-involved thrusting and uplift of the Apuseni Mountains. This was coeval with the late stage gravitational spreading of the post-salt succession.

When compared with other intra-Carpathian basins the Transylvanian Basin evolution differs during the Middle to Late Miocene (e.g. Krézsek and Filipescu, 2005) due to its proximity to the deep lithospheric processes in the Carpathians subduction zone (e.g. Cloetingh et al., 2004). The main differences include the lack of extensional structures and not related to salt tectonics, high post-salt subsidence rates followed by Pliocene uplift and erosion.

Balanced cross sections have been constructed to get a feeling for the magnitudes of displacement in the foreland fold belt and to constrain the Carpathians uplift, which played a major role in the gravitational spreading. The inferred southwestward dipping basement slope was induced by the uplift of the central-northern Carpathians, as indicated by fission track results (e.g. Sanders, 1999) and lithospheric modeling (e.g. Cloetingh et al., 2004). The timing of the Carpathians uplift, backarc volcanism and gravitational spreading of the Transylvanian Basin fits well.

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FIGURE CAPTIONS

Fig. 1. Overview of Alps, Carpathians and Dinarides. The grey areas are the most important Paratethys basin systems: Alpine (Western Paratethys), Intra-Carpathian (Central Paratethys) and Dacian, Euxinian and Caspian (Eastern Paratethys). The Transylvanian (TB) and Pannonian basins (PB) belong to the Intra-Carpathian basin system.

Fig. 2. Pre-Cenomanian reconstructions (modified after Săndulescu, 1988).

Fig. 3. Basement map of Transylvanian Basin including all well control.

Fig 4. Seismic closeup shows seismic tie to the 6042Deleni well. For the stratigraphy, see discussion in the text. The age of the normal fault is post-Aptian, because it cuts the Lower Cretaceous and is sealed by Upper Cretaceous. This and other faults were active during the early stages of rifting in the Târnave Basin. For the location, refer to Fig. 6.

Fig. 5. Stratigraphic chart and evolution summary of the Transylvanian Basin with key formation names based on Koch (1894, 1900), Ciupagea et al. (1970), Mészáros et al. (1971), Dicea et al. (1980a), Lupu and Lupu (1983), Petrescu et al., (1987, 1989); Rusu (1989, 1995), Mészáros (1991, 2000), Nicorici and Mészáros (1994), Chira (1994), Popescu et al. (1995), Filipescu (1996, 2001), Proust and Hosu (1996), De Broucker et al. (1998), Codrea and Hosu (2001), Schuller (2004), Krézsek (2005), Kaminski and Filipescu (2005), Krézsek and Filipescu (2005), and this paper. A simplified lithological column and tectonic regimes of the Eastern Carpathians based on works of Săndulescu (1988), Mațenco (1997), Linzer et al. (1998), Zweigel et al. (1998), and Tărăpoancă et al. (2003, 2004) is shown also. Standard absolute ages are from Gradstein et al. (2004). Paratethys stage boundaries follow Rögl (1996), Filipescu (2001) and Krézsek and Filipescu (2005).

Fig. 6. Geological sketch map of the Transylvanian Basin and surroundings. Shows also the grid of regional seismic lines and close-ups selected for paper and the location of regional transects through the Eastern Carpathians (Plate 7).

Fig. 7. Pre-Tertiary subcrop map showing Upper Cretaceous subsurface thickness, significant normal faults, Upper Cretaceous well penetrations and outcrop distribution. Two of most important Upper Cretaceous basins, i.e. the Puini and Târnave basins, are presented as closeups.

Fig. 8. Uninterpreted and interpreted seismic closeup of Plate 4, showing details of the Upper Cretaceous Târnave Basin. For the location, refer to Fig. 6.

Fig. 9. Uninterpreted and interpreted seismic closeup of Plate 1, showing details of Santonian to Maastrichtian sag fill of the Puini Basin, which seals the Puini thrust front and its Albian to Coniacian flexural fill. Thus, the Latest Cretaceous subsidence in the Puini basin was not related to the Puini thrust front. For the location, refer to Fig. 6.

Fig. 10. Cartoon showing reconstructed crystalline wedge top Late Cretaceous basins (Dej and Puini), Mid Cretaceous Carpathians folded belt and foredeep facing the distal part of the East European passive margin. The section is located on Fig. 7.

Fig. 11. Uninterpreted and interpreted seismic closeup of Plate 1, showing relation of Paleocene/Eocene to Upper Cretaceous in the Puini basin area and Sag1-2-3 fill. For the location, refer to Fig. 6.

Fig. 12. Uninterpreted and interpreted seismic closeup of Plate 2, showing relation of Paleocene/Eocene to Upper Cretaceous in the Puini basin area and Sag1-2-3 fill. For the location, refer to Fig. 6.

Fig. 13. Pre-Oligocene subcrop map showing the Paleocene/Eocene outcrop, facies and subsurface thickness.

Fig. 14. Pre-Miocene subcrop map showing Oligocene outcrop, facies and subsurface thickness.

Fig. 15. Regional cross section of Transylvanian Basin to Carpathians foredeep, showing reconstructed Oligocene paleogeography. Location marked on Fig. 14.

Fig. 16. Detail of Plate 5, flattened on base Middle Miocene, to show onlap of the Lower Miocene in the distal reaches (i.e. the central part of the Transylvanian Basin) of the Lower Miocene flexural basin.

Fig. 17. Pre-Neogene subcrop map showing Lower Miocene outcrops, subsurface thickness distribution and lithofacies. Paleogeographical reconstruction for Eggenburgian.

Fig. 18. Early Miocene regional cross-section of Pienides, Transylvanian Flexural Basin and Carpathian fold belt and its foreland. Location marked on Fig. 17.

Fig. 19. Pre-Middle Miocene subcrop map. Shows subsurface salt thickness distribution, salt outcrops and facies distribution. The areas shown as “nondeposition/erosion” represented uplifted areas during salt deposition.

Fig. 20. Post-Salt thickness distribution (i.e. truncated isopach) with Upper Badenian onlap band in grey. Paleogeographical reconstruction for the Early Sarmatian.

Fig. 21. Regional salt paleogeography (Mid Badenian) showing a narrow ridge of shallow accretionary wedge and direct connection to foreland basin salt, as well erosion/continental deposition in the Paleo Apuseni Mountains. Location marked on Fig. 19.

Fig. 22. Regional post-salt Late Badenian paleogeography, showing a submerged Carpathian fold belt and direct connection to the foreland basin. Location marked on Fig. 20.

Fig. 23. Pliocene cross-section. The Transylvanian Basin is eroded, while sedimentation continue in Carpathians wedge top basins (e.g. Braşov basin) and its foreland. Volcanism is active on the margins of the Transylvanian Basin. The location is the same as Fig. 22.

Fig. 24. Close-up map of the Transylvanian Basin that illustrates the major salt tectonic features, including post-salt anticlines and synclines, toe-thrusts, over thrusts, compressional lateral ramps, inferred detachment scarp under

the volcanics, salt pillows (thicker then 200 ms), salt walls and outcropping salt diapirs. Fission track uplift ages of Sanders (1999) are indicated by numbers in boxes and the ages of volcanism (Pécskay et al., 1995) by encircled numbers. See discussion in text.

Fig. 25. Geological map of Gurghiu area and surface and dips profiles updated from Bucur et al. 1971. The sections are non-exaggerated. The area location is shown on Fig. 24.

Fig. 26 Seismic detail interpreted and non-interpreted in time that show extensional growth fault and salt weld in the Eastern Transylvanian Basin. Location marked on Fig. 6.

Fig. 27. Seismic detail interpreted and non-interpreted in time that shows extensional growth fault and salt weld in the Eastern Transylvanian Basin. The seismic line has been extended eastward using outcrop data in order to show the outcropping detachment scarp north of the volcanic pile. Location marked Fig. 6.

Fig. 28. Example of lateral ramp to the south (Ruşi-Cenade fault). Location marked on Fig. 6.

PLATES

PLATE 1. Uninterpreted and interpreted drawing of east-west oriented regional section. The depth is in time (sec). For the location, please refer to Fig. 6. Legend is on Plate 2.

PLATE 2. Uninterpreted and interpreted drawing of east-west oriented regional section. Below the interpreted drawing a simplified version of the nonexagerated transect is shown also. The depth is in time (sec). For the location, refer to Fig. 6. The transect shows the relation between the Transylvanian Basin and Apuseni Mts./Inner Eastern Carpathians, Upper Cretaceous to Eocene basins on the top of Middle Cretaceous Transylvanian suture, salt structure domains, architecture of the post-salt succession, and evidence for late stage (Pliocene) gravitational spreading.

PLATE 3. Uninterpreted and interpreted drawing of east-west oriented regional section. The depth is in time (sec). For the location, refer to Fig. 6. Legend is on Plate 2. For discussion, see in the text.

PLATE 4. Uninterpreted and interpreted drawing of east west oriented regional section. The depth is in time (sec). For the location, refer to Fig. 6. Legend is on Plate 2. For discussion, see in the text.

PLATE 5. Uninterpreted and interpreted drawing of north south oriented regional section. The depth is in time (sec). For the location, refer to Fig. 6. Legend is on Plate 2. For discussion, see in the text.

PLATE 6. Uninterpreted and interpreted drawing of north south oriented regional section. The depth is in time (sec). For the location, refer to Fig. 6. Legend is on Plate 2. For discussion, see in the text.

PLATE 7. Two balanced cross sections (Transect 1 and 2) and reconstructions. For the location, please refer to Fig. 6.

TABLES

Table 1. List of wells and type of basement penetrated.

Fig. 1

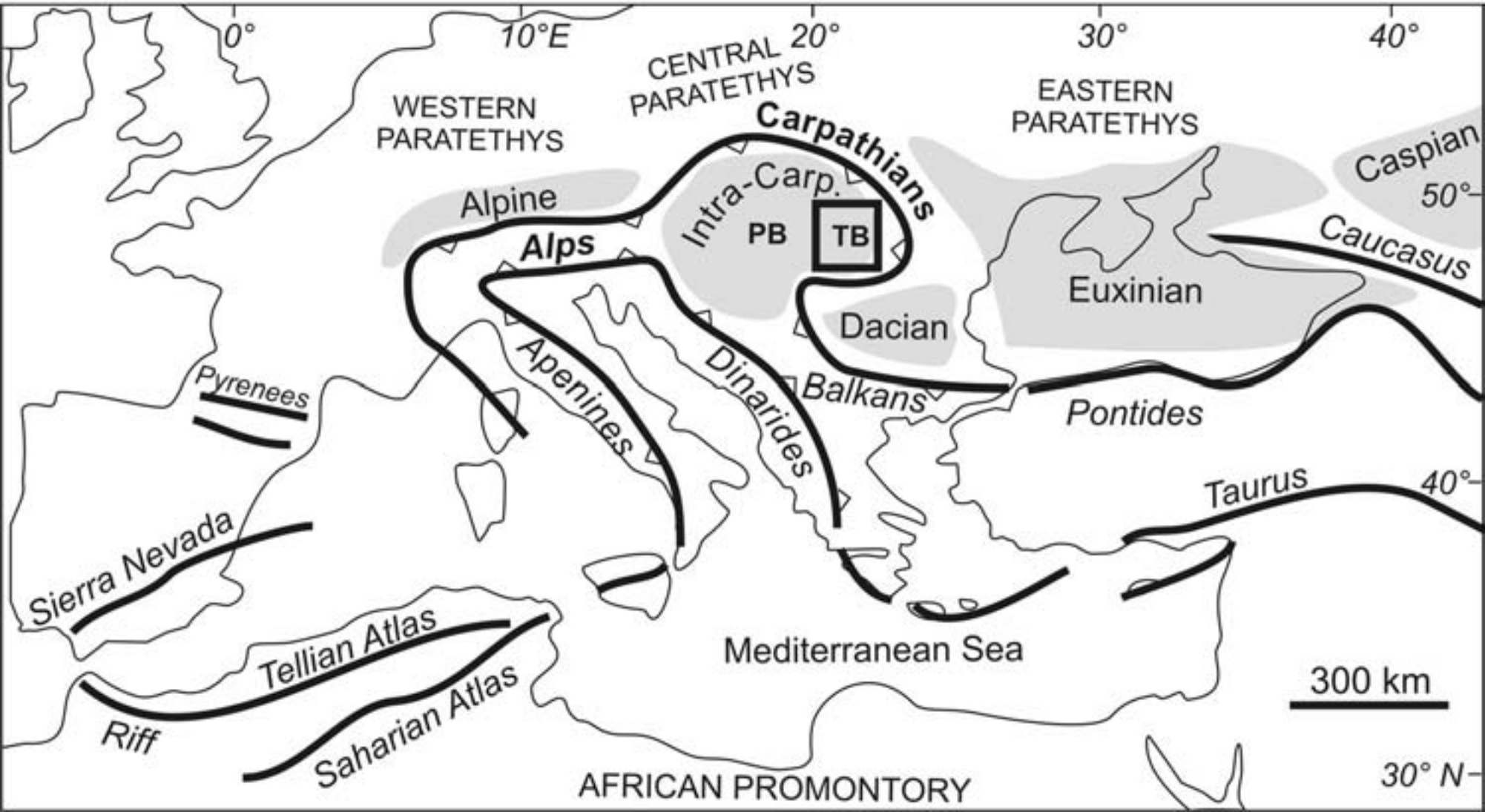

0°
10°E
20°
30°
40°
CENTRAL PARATETHYS
WESTERN PARATETHYS
EASTERN PARATETHYS
Carpathians
Intra-Carp.
PB
TB
Alpine
Alps
Caspian
50°
Caucasus
Euxinian
Dacian
Apenines
Dinarides
Balkans
Pontides
Pyrenees
Taurus
40°
Sierra Nevada
Tellian Atlas
Mediterranean Sea
300 km
Riff
Saharian Atlas
AFRICAN PROMONTORY
30° N

Fig. 2

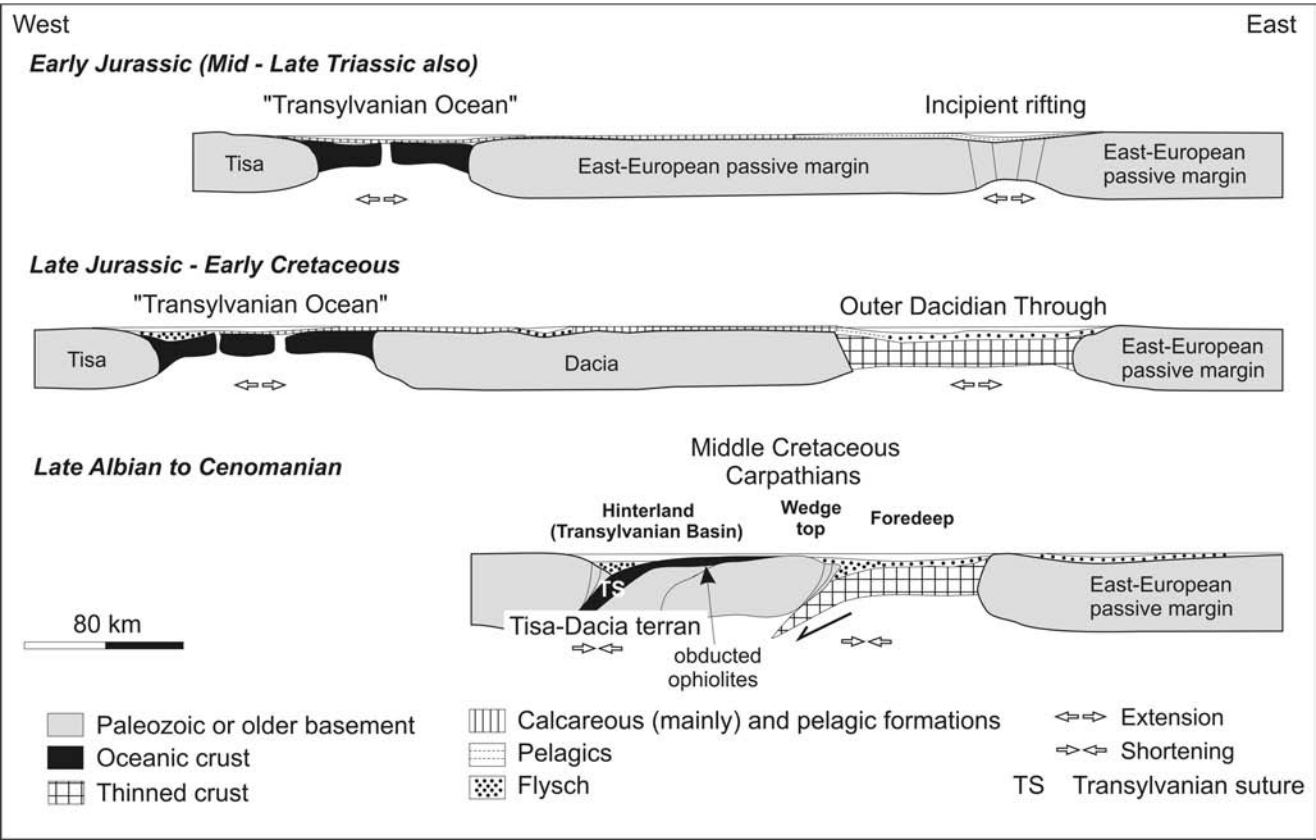

West
East
Early Jurassic (Mid - Late Triassic also)
"Transylvanian Ocean"
Incipient rifting
Tisa
East-European passive margin
East-European passive margin
Late Jurassic - Early Cretaceous
"Transylvanian Ocean"
Outer Dacidian Through
Tisa
Dacia
East-European passive margin
Late Albian to Cenomanian
Middle Cretaceous Carpathians
Hinterland (Transylvanian Basin)
Wedge top
Foredeep
TS
East-European passive margin
Tisa-Dacia terran
obducted ophiolites
80 km
Paleozoic or older basement
Oceanic crust
Thinned crust
Calcareous (mainly) and pelagic formations
Pelagics
Flysch
Extension
Shortening
TS Transylvanian suture

Fig. 3

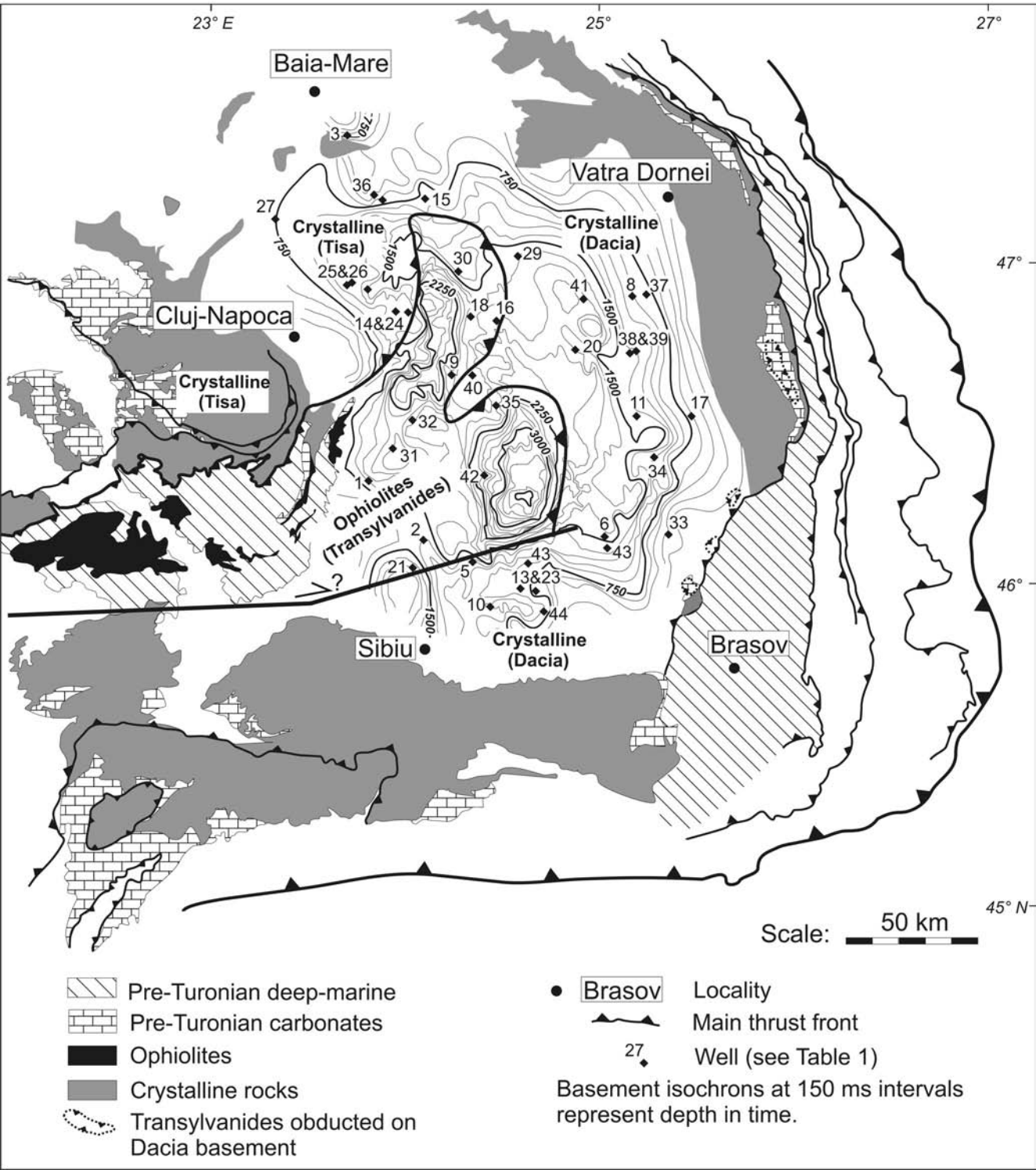

23° E
25°
27°
47°
46°
45° N
Baia-Mare
Vatra Dornei
Cluj-Napoca
Sibiu
Brasov
Crystalline (Tisa)
Crystalline (Dacia)
Crystalline (Tisa)
Crystalline (Dacia)
Ophiolites (Transylvanides)
?
750
1500
2250
3000
Scale: 50 km
Pre-Turonian deep-marine
Pre-Turonian carbonates
Ophiolites
Crystalline rocks
Transylvanides obducted on Dacia basement
Brasov Locality
Main thrust front
Well (see Table 1)
Basement isochrons at 150 ms intervals represent depth in time.

Fig. 4

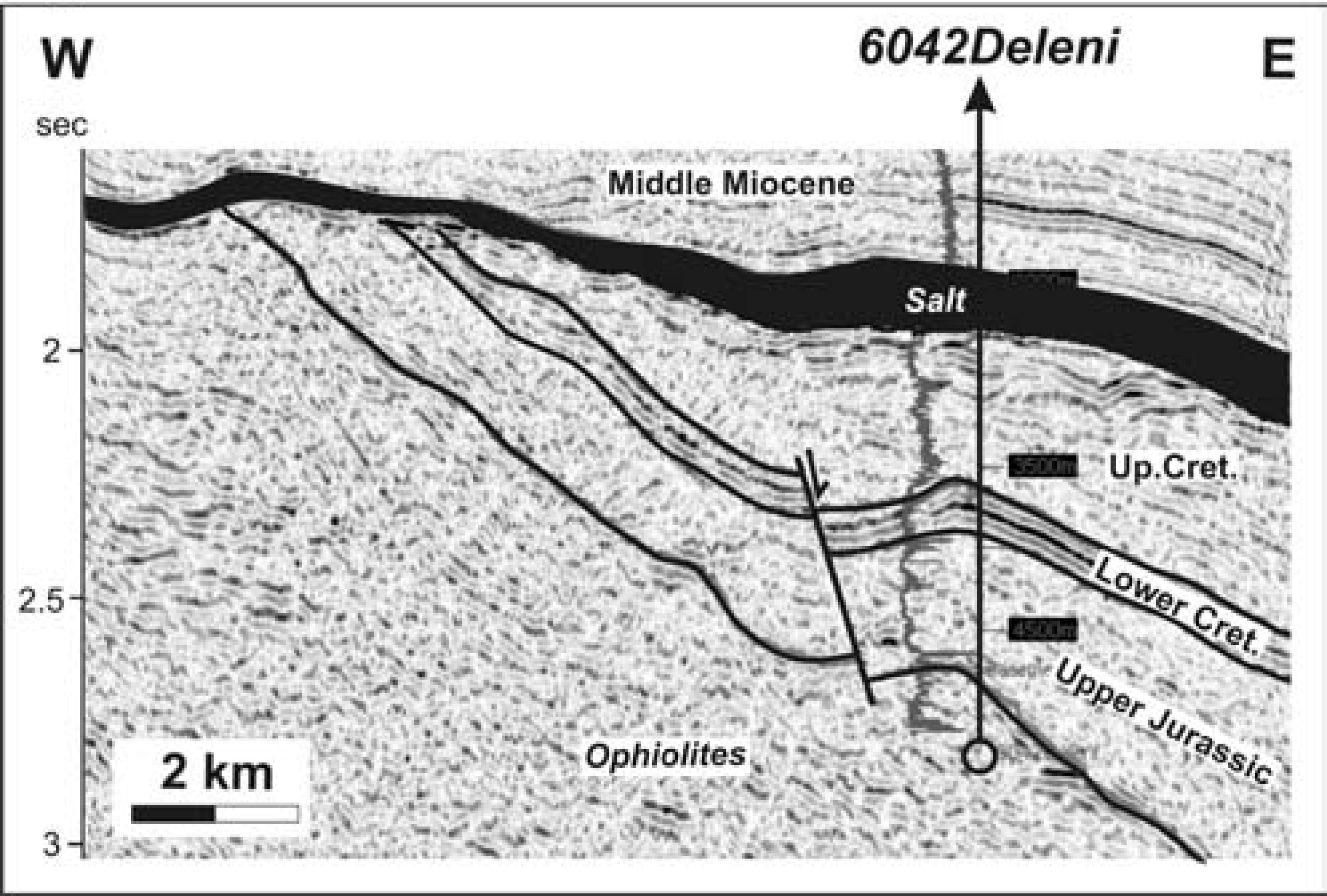

W
E
6042Deleni
sec
2
2.5
3
Middle Miocene
Salt
Up.Cret.
Lower Cret.
Upper Jurassic
Ophiolites
2 km

Fig. 5

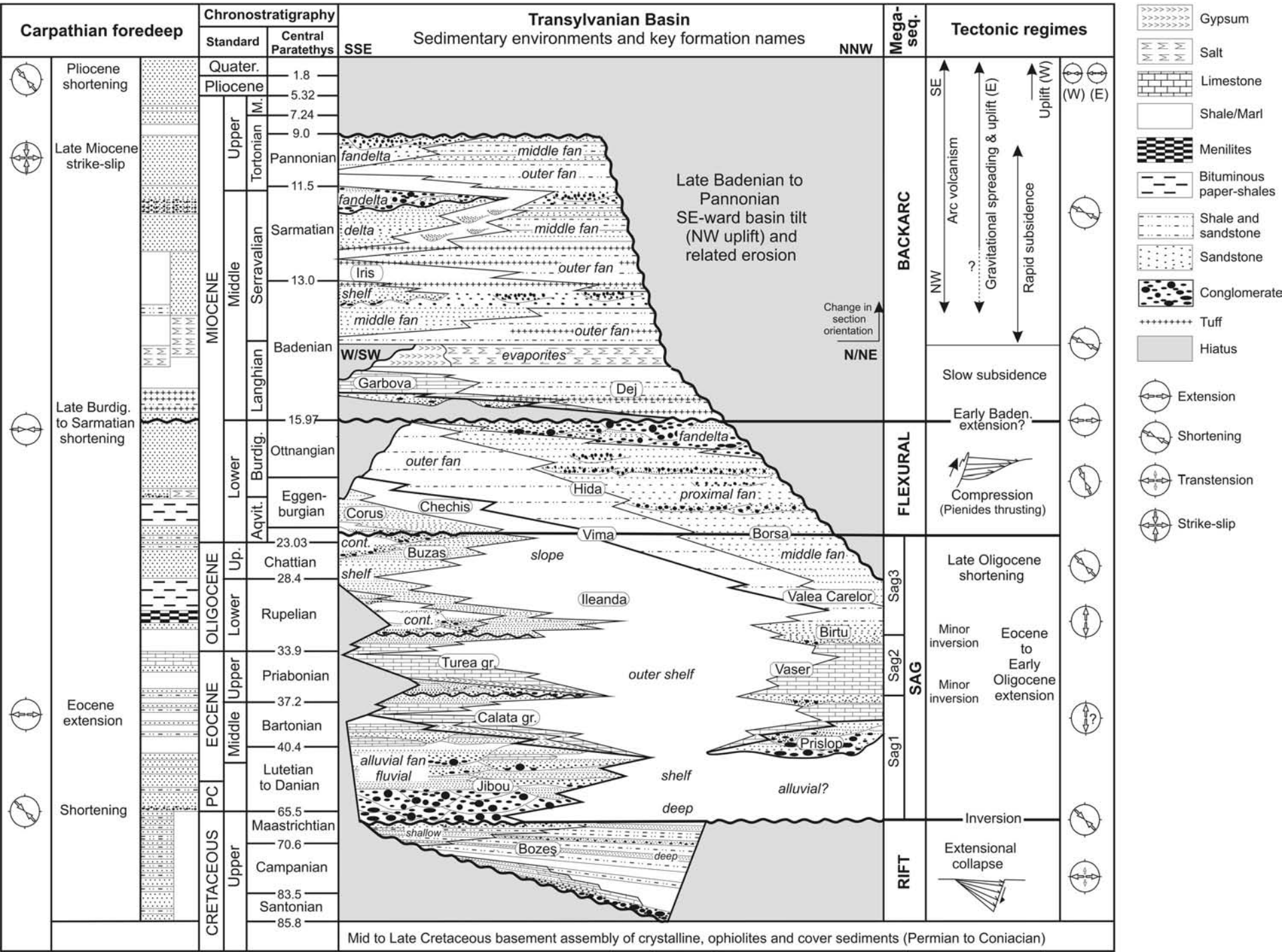

Carpathian foredeep
Chronostratigraphy
Standard
Central Paratethys
Transylvanian Basin
Sedimentary environments and key formation names
SSE
NNW
Mega-seq.
Tectonic regimes
Pliocene shortening
Late Miocene strike-slip
Late Burdig. to Sarmatian shortening
Eocene extension
Shortening
Quater.
Pliocene
MIOCENE
Upper
Middle
Lower
Tortonian
Serravalian
Langhian
Burdig.
Aqvit.
OLIGOCENE
Up.
Lower
EOCENE
Upper
Middle
PC
CRETACEOUS
Upper
1.8
5.32
7.24
9.0
11.5
13.0
15.97
23.03
28.4
33.9
37.2
40.4
65.5
70.6
83.5
85.8
Pannonian
Sarmatian
Badenian
Ottnangian
Eggen-burgian
Chattian
Rupelian
Priabonian
Bartonian
Lutetian to Danian
Maastrichtian
Campanian
Santonian
fandelta
middle fan
outer fan
delta
Iris
shelf
evaporites
W/SW
Garbova
Dej
Late Badenian to Pannonian SE-ward basin tilt (NW uplift) and related erosion
Change in section orientation
N/NE
Hida
proximal fan
Chechis
Corus
Vima
Borsa
cont.
Buzas
slope
Ileanda
Valea Carelor
Birtu
Turea gr.
outer shelf
Vaser
Calata gr.
Prislop
alluvial fan
fluvial
Jibou
alluvial?
deep
shallow
Bozeș
Mid to Late Cretaceous basement assembly of crystalline, ophiolites and cover sediments (Permian to Coniacian)
BACKARC
FLEXURAL
SAG
RIFT
Sag3
Sag2
Sag1
SE
NW
Arc volcanism
Gravitational spreading & uplift (E)
Uplift (W)
Rapid subsidence
Slow subsidence
Early Baden. extension?
Compression (Pienides thrusting)
Late Oligocene shortening
Minor inversion
Eocene to Early Oligocene extension
Inversion
Extensional collapse
(W) (E)
Gypsum
Salt
Limestone
Shale/Marl
Menilites
Bituminous paper-shales
Shale and sandstone
Sandstone
Conglomerate
Tuff
Hiatus
Extension
Shortening
Transtension
Strike-slip

Fig. 6

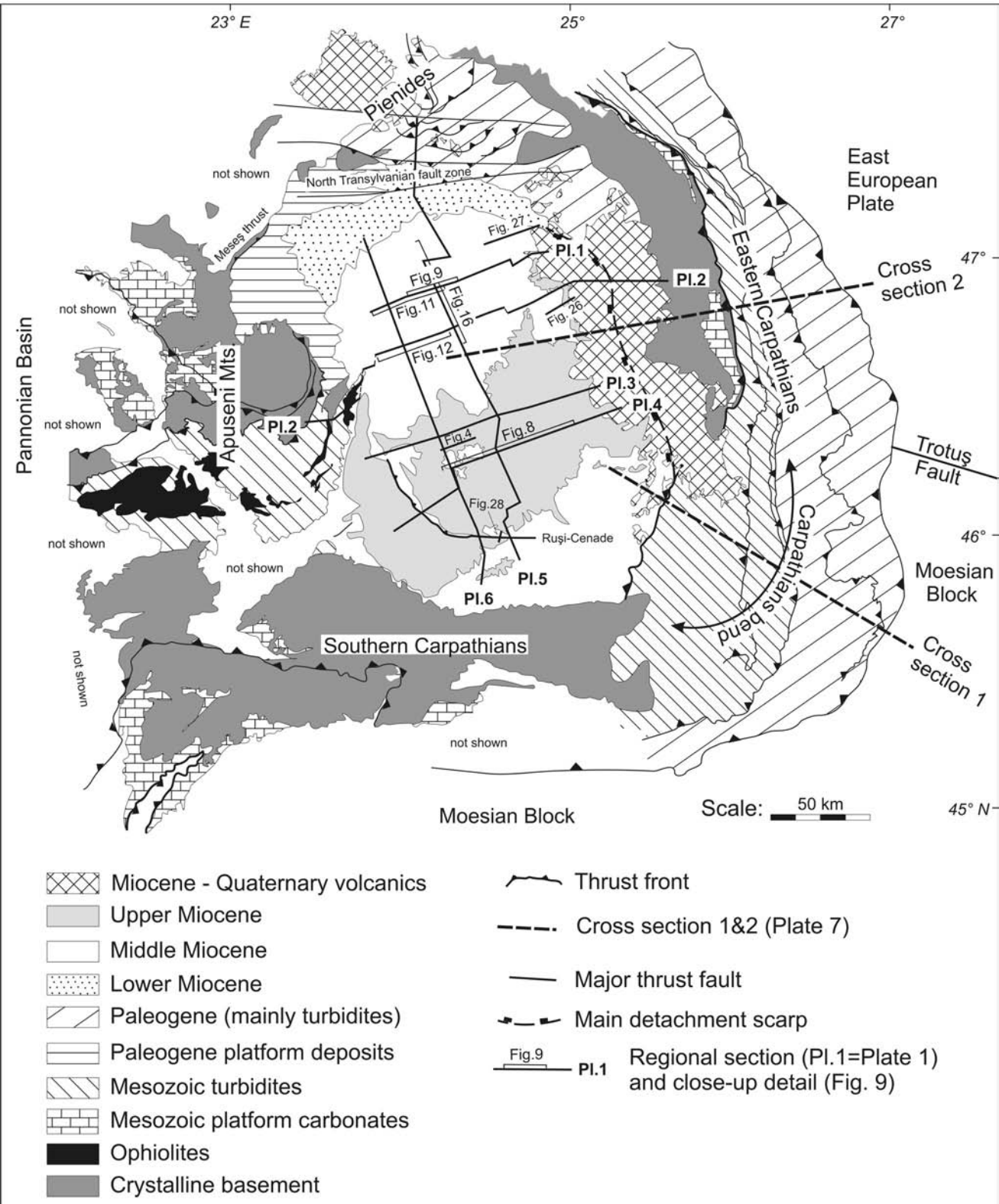

23° E
25°
27°
47°
46°
45° N
Pienides
North Transylvanian fault zone
not shown
Meseş thrust
East
European
Plate
Eastern Carpathians
Cross
section 2
Pl.1
Pl.2
Fig. 27
Fig. 9
Fig. 11
Fig. 16
Fig. 26
Fig. 12
Pl.3
Pl.4
Pannonian Basin
Apuseni Mts
Fig. 4
Fig. 8
Trotuş
Fault
Fig. 28
Ruşi-Cenade
Carpathians bend
Moesian
Block
Pl.5
Pl.6
Southern Carpathians
Cross
section 1
Moesian Block
Scale:
50 km
Miocene - Quaternary volcanics
Upper Miocene
Middle Miocene
Lower Miocene
Paleogene (mainly turbidites)
Paleogene platform deposits
Mesozoic turbidites
Mesozoic platform carbonates
Ophiolites
Crystalline basement
Thrust front
Cross section 1&2 (Plate 7)
Major thrust fault
Main detachment scarp
Fig.9
Pl.1
Regional section (Pl.1=Plate 1)
and close-up detail (Fig. 9)

Fig. 7

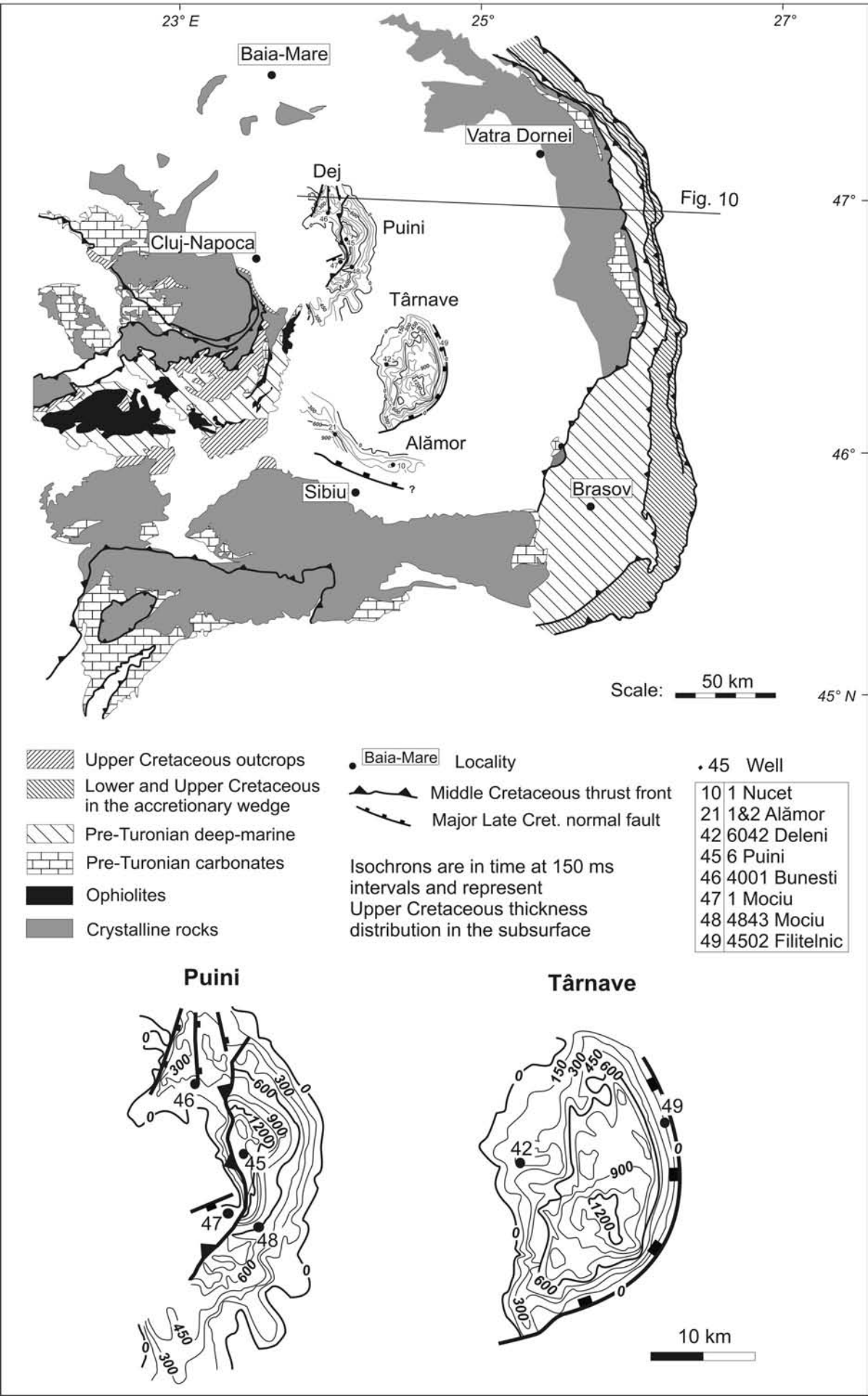

23° E
25°
27°
47°
46°
45° N
Baia-Mare
Vatra Dornei
Dej
Fig. 10
Puini
Cluj-Napoca
Târnave
Alămor
Sibiu
Brasov
Scale:
50 km
Upper Cretaceous outcrops
Lower and Upper Cretaceous in the accretionary wedge
Pre-Turonian deep-marine
Pre-Turonian carbonates
Ophiolites
Crystalline rocks
Baia-Mare Locality
Middle Cretaceous thrust front
Major Late Cret. normal fault
Isochrons are in time at 150 ms intervals and represent Upper Cretaceous thickness distribution in the subsurface
45 Well
10 1 Nucet
21 1&2 Alămor
42 6042 Deleni
45 6 Puini
46 4001 Bunesti
47 1 Mociu
48 4843 Mociu
49 4502 Filitelnic
Puini
Târnave
10 km

Fig. 8

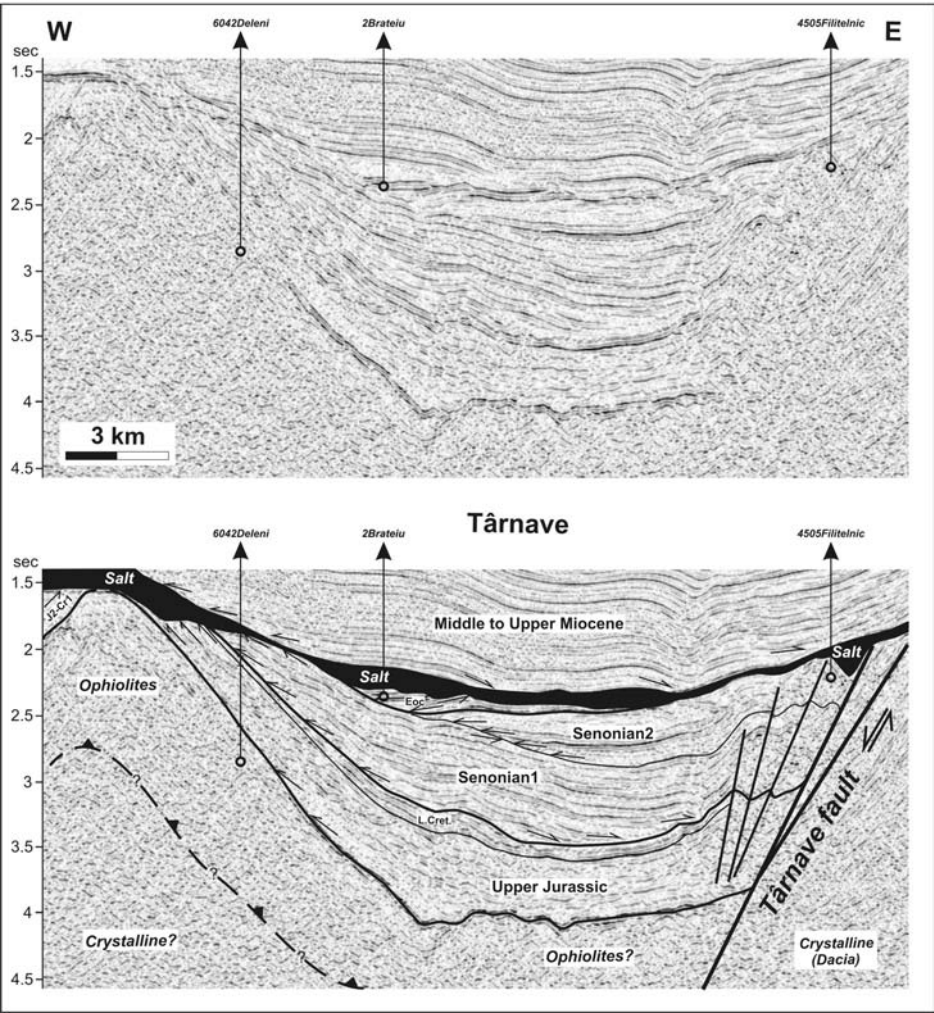

W
E
6042Deleni
2Brateiu
4505Filitelnic
sec
1.5
2
2.5
3
3.5
4
4.5
3 km
Târnave
Salt
J2-Cr1
Middle to Upper Miocene
Ophiolites
Eoc
Senonian2
Senonian1
L.Cret
Upper Jurassic
Târnave fault
Crystalline?
Ophiolites?
Crystalline
(Dacia)

Fig. 9

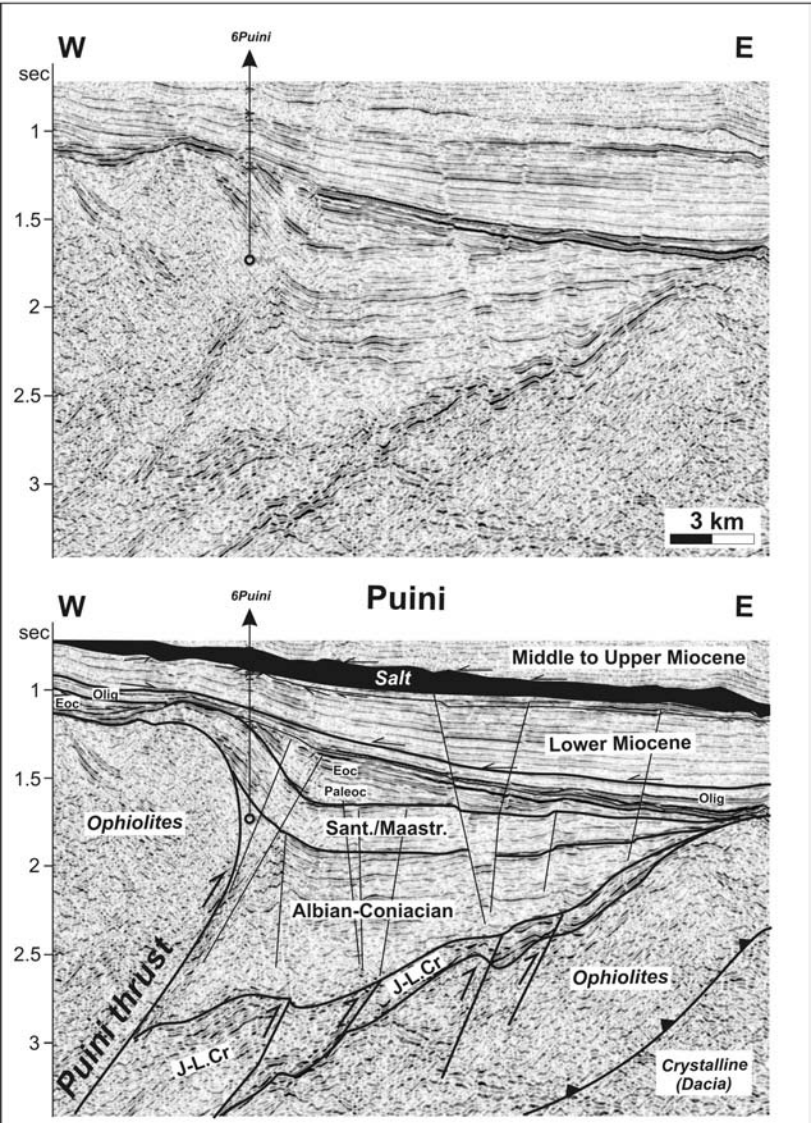

W
E
6Puini
sec
1
1.5
2
2.5
3
3 km
Puini
Middle to Upper Miocene
Salt
Olig
Eoc
Lower Miocene
Eoc
Paleoc
Olig
Ophiolites
Sant./Maastr.
Albian-Coniacian
J-L Cr
Ophiolites
Puini thrust
J-L Cr
Crystalline
(Dacia)

Fig. 10

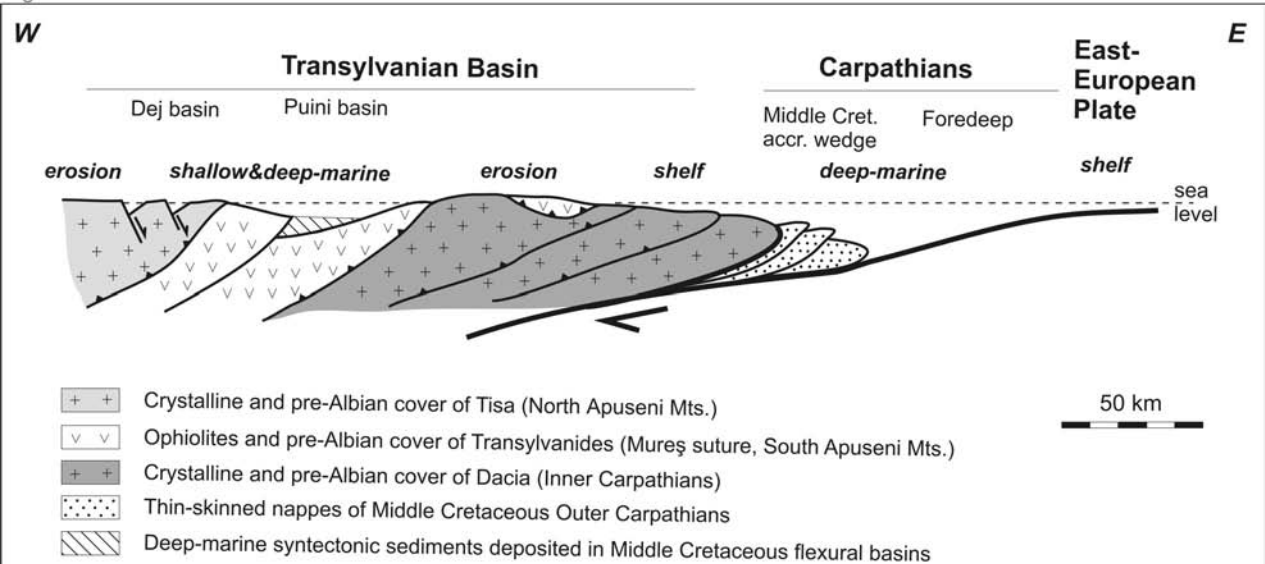

W
E
Transylvanian Basin
Carpathians
East-European Plate
Dej basin
Puini basin
Middle Cret. accr. wedge
Foredeep
erosion
shallow&deep-marine
erosion
shelf
deep-marine
shelf
sea level
Crystalline and pre-Albian cover of Tisa (North Apuseni Mts.)
Ophiolites and pre-Albian cover of Transylvanides (Mureş suture, South Apuseni Mts.)
Crystalline and pre-Albian cover of Dacia (Inner Carpathians)
Thin-skinned nappes of Middle Cretaceous Outer Carpathians
Deep-marine syntectonic sediments deposited in Middle Cretaceous flexural basins
50 km

Fig. 11

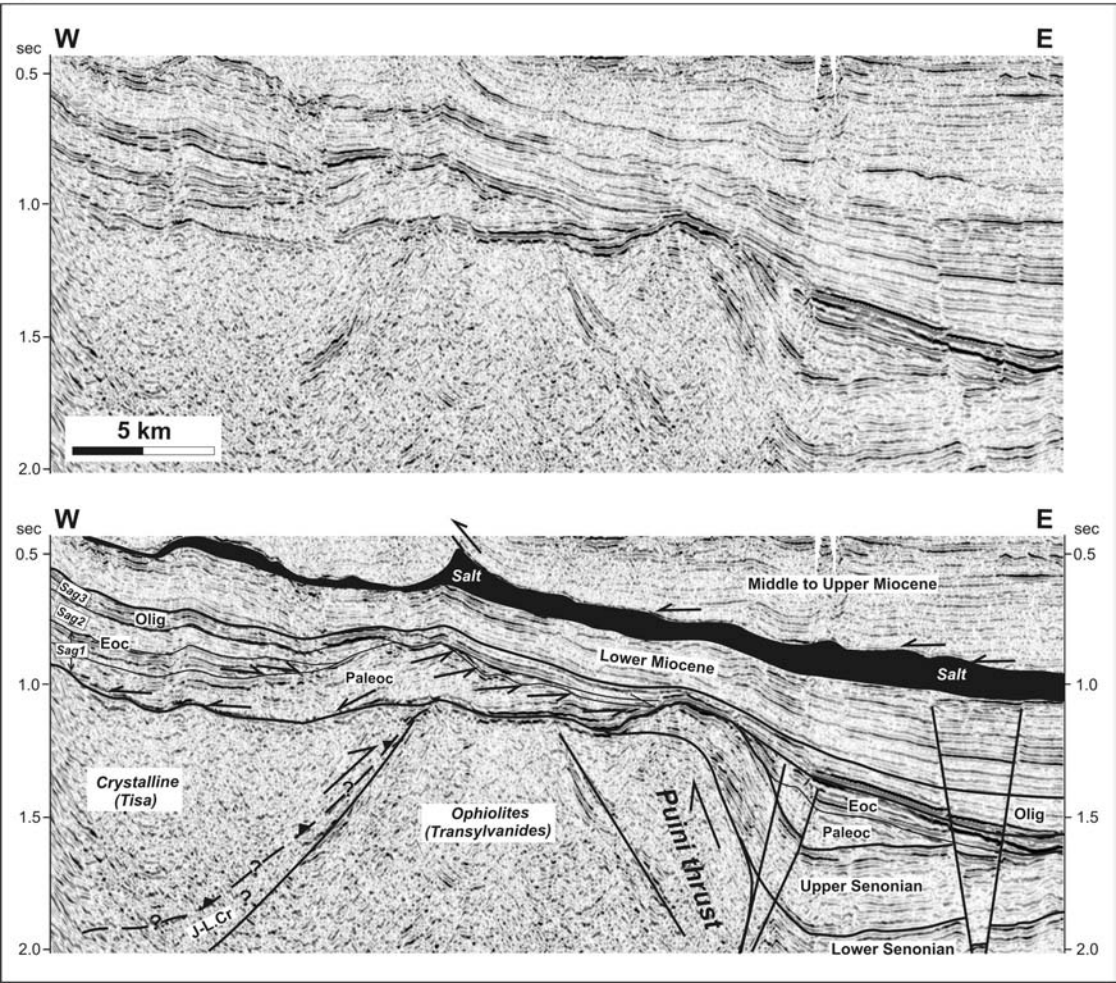

W
E
sec
0.5
1.0
1.5
2.0
5 km
W
E
sec
0.5
1.0
1.5
2.0
Salt
Middle to Upper Miocene
Sag3
Sag2
Sag1
Olig
Eoc
Lower Miocene
Salt
Paleoc
Crystalline
(Tisa)
Ophiolites
(Transylvanides)
Puini thrust
Eoc
Paleoc
Olig
Upper Senonian
J-L.Cr
Lower Senonian

Fig. 12

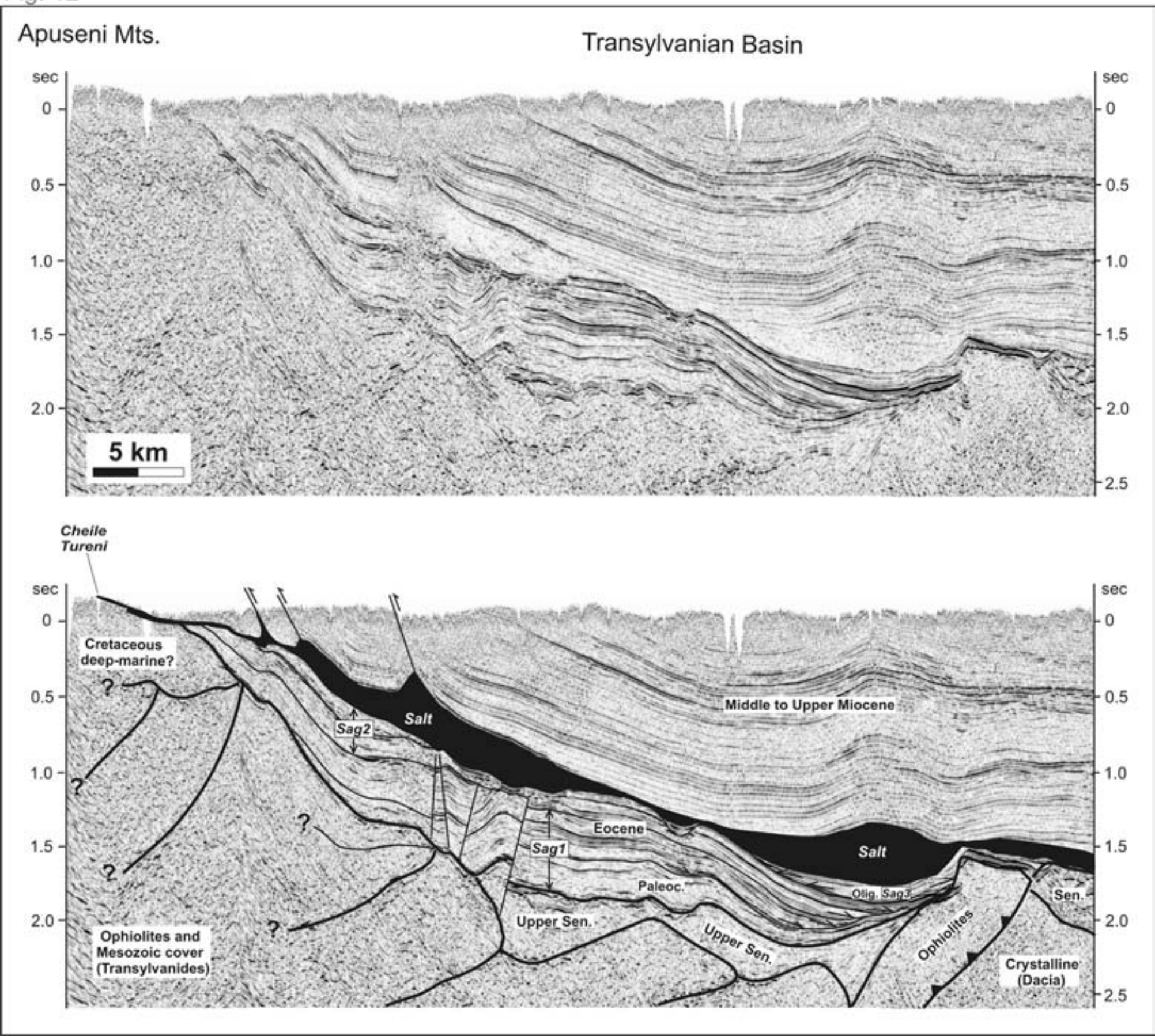
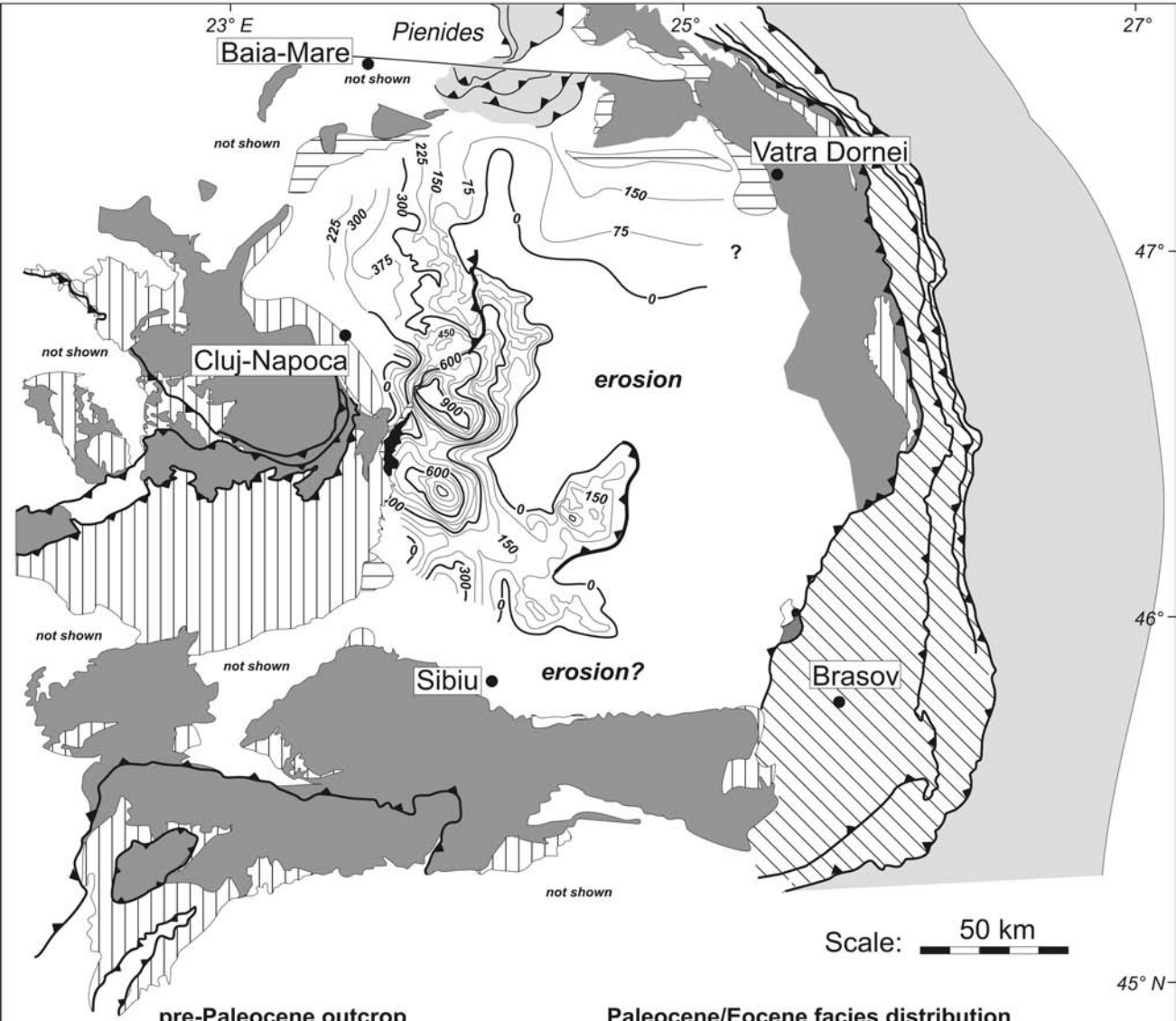

Apuseni Mts.
Transylvanian Basin
sec
0
0.5
1.0
1.5
2.0
2.5
5 km
Cheile Tureni
Cretaceous deep-marine?
Salt
Sag2
Middle to Upper Miocene
Eocene
Sag1
Salt
Paleoc.
Olig. Sag3
Sen.
Upper Sen.
Upper Sen.
Ophiolites
Ophiolites and Mesozoic cover (Transylvanides)
Crystalline (Dacia)

Fig.13



pre-Paleocene outcrop

- pre-Paleocene sedimentary cover
- Mesozoic deep-marine in the Carpathians accretionary wedge
- Crystalline basement

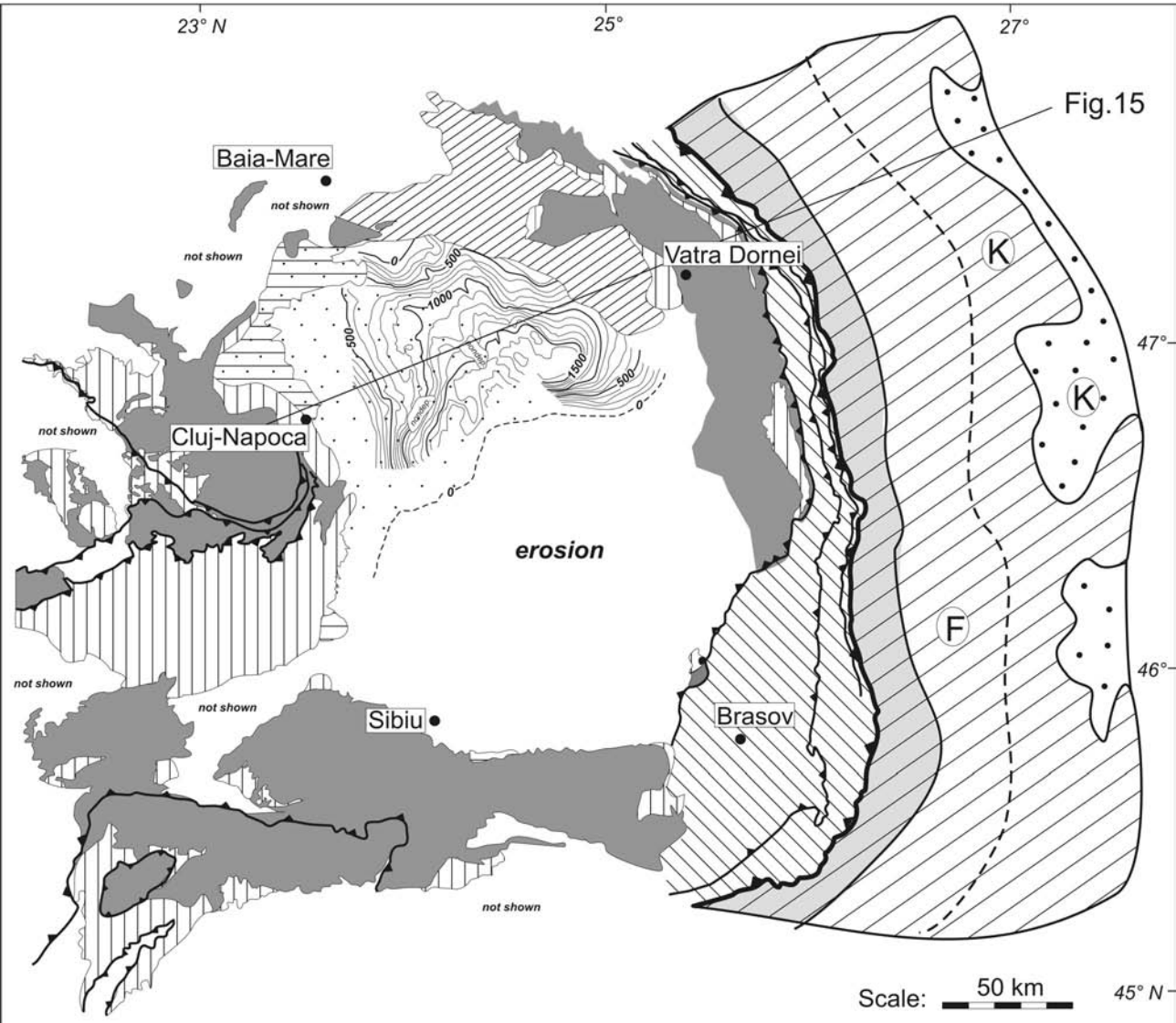
- Baia-Mare — Locality
- Main thrust front
- Major normal fault

Paleocene/Eocene facies distribution

- Continental to shallow-marine outcrop
- Deep-marine outcrop
- Continental to shallow-marine

Isochrons are in time at 75 ms intervals and represent Paleocene-Eocene continental to shallow-marine thickness distribution in the subsurface.

Fig.14



pre-Oligocene outcrop

- pre-Oligocene sedimentary cover
- Mesozoic deep-marine in the Carpathians Cretaceous accretionary wedge
- Crystalline basement
- Baia-Mare Locality
- Front of the Cretaceous Carpathians

Oligocene outcrop and facies distribution

Transylvanian Basin

- Continental to inner-shelf outcrop
- Outer-shelf to deep-marine outcrop
- Continental to inner-shelf
- Outer-shelf to slope

Isochrons are in time at 100 ms intervals and represent Oligocene thickness distribution in the subsurface.

Eastern Carpathian foredeep (Rupelian)

- Fan delta and inner-shelf
- Slope
- Deep-marine

F Fusaru K Kliwa

Fig. 15

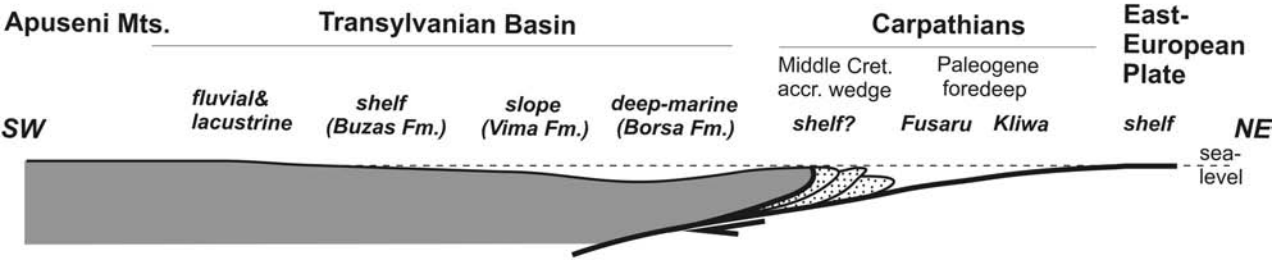

Apuseni Mts.
Transylvanian Basin
Carpathians
East-European Plate
Middle Cret. accr. wedge
Paleogene foredeep
SW
fluvial& lacustrine
shelf (Buzas Fm.)
slope (Vima Fm.)
deep-marine (Borsa Fm.)
shelf?
Fusaru
Kliwa
shelf
NE
sea-level
Basement and pre-Paleogene cover of TISA-DACIA
Middle Cretaceous fold belt (Jurassic - Middle Cretaceous)
Scale:
50 km

Fig. 16

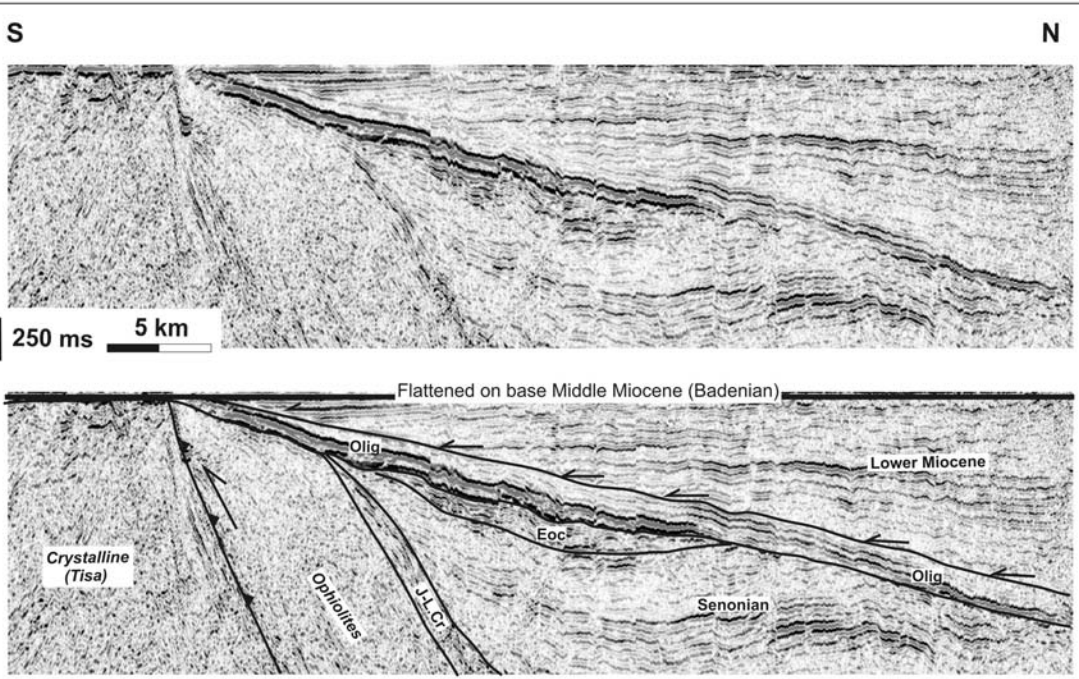
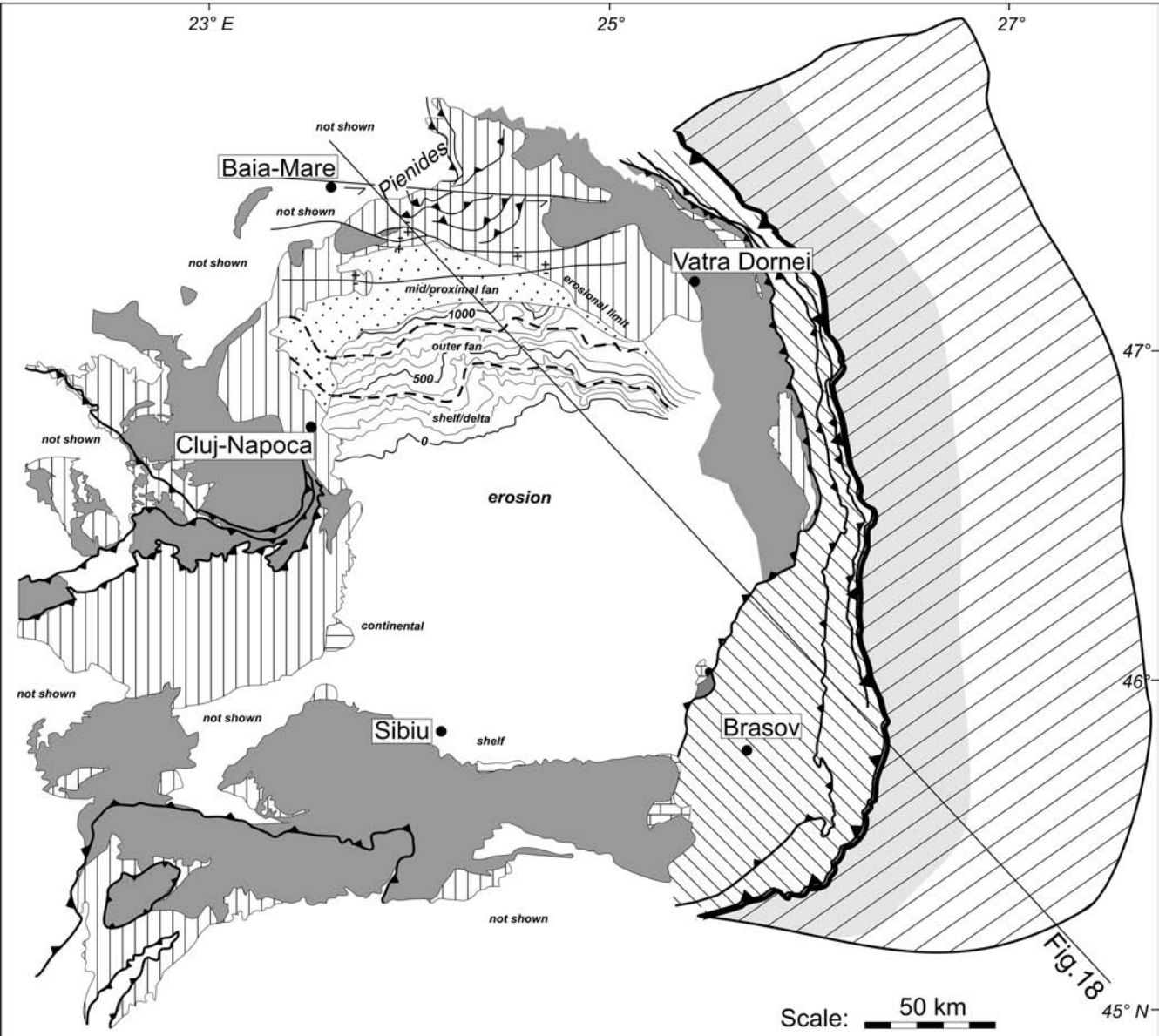

S
N
250 ms
5 km
Flattened on base Middle Miocene (Badenian)
Olig
Lower Miocene
Eoc
Olig
Senonian
Crystalline
(Tisa)
Ophiolites
J-L Cr

Fig. 17



pre-Miocene outcrop

- pre-Miocene sedimentary cover
- Mesozoic deep-marine in the Carpathians accretionary wedge
- Crystalline basement

- Baia-Mare — Locality
- Main thrust front

Lower Miocene outcrop and facies distribution

Transylvanian Basin

- Outcrop
- Subcrop. Isochrons are in time at 100 ms intervals and represent Early Miocene thickness distribution

Eastern Carpathian foredeep (Eggenburgian reconstruction)

- Salt
- Gypsum
- Facies boundary

Fig. 18

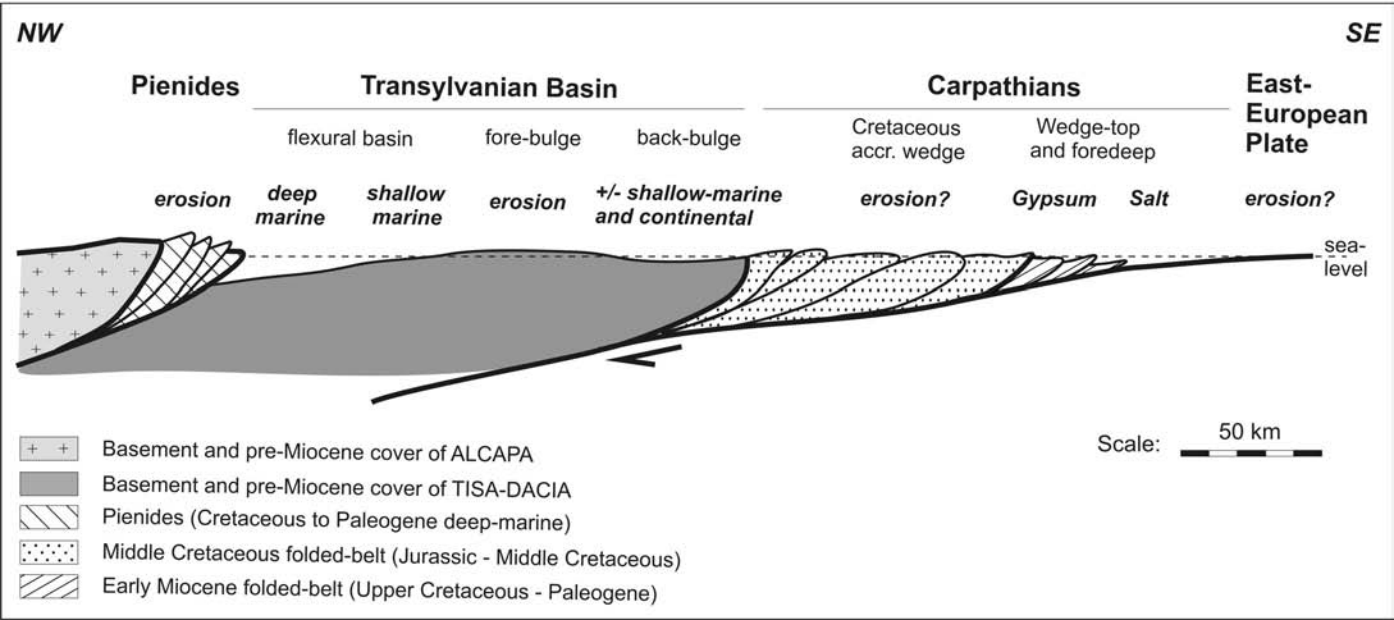

NW
SE
Pienides
Transylvanian Basin
Carpathians
East-European Plate
flexural basin
fore-bulge
back-bulge
Cretaceous accr. wedge
Wedge-top and foredeep
erosion
deep marine
shallow marine
erosion
+/- shallow-marine and continental
erosion?
Gypsum
Salt
erosion?
sea-level
Scale:
50 km
Basement and pre-Miocene cover of ALCAPA
Basement and pre-Miocene cover of TISA-DACIA
Pienides (Cretaceous to Paleogene deep-marine)
Middle Cretaceous folded-belt (Jurassic - Middle Cretaceous)
Early Miocene folded-belt (Upper Cretaceous - Paleogene)

Fig. 19

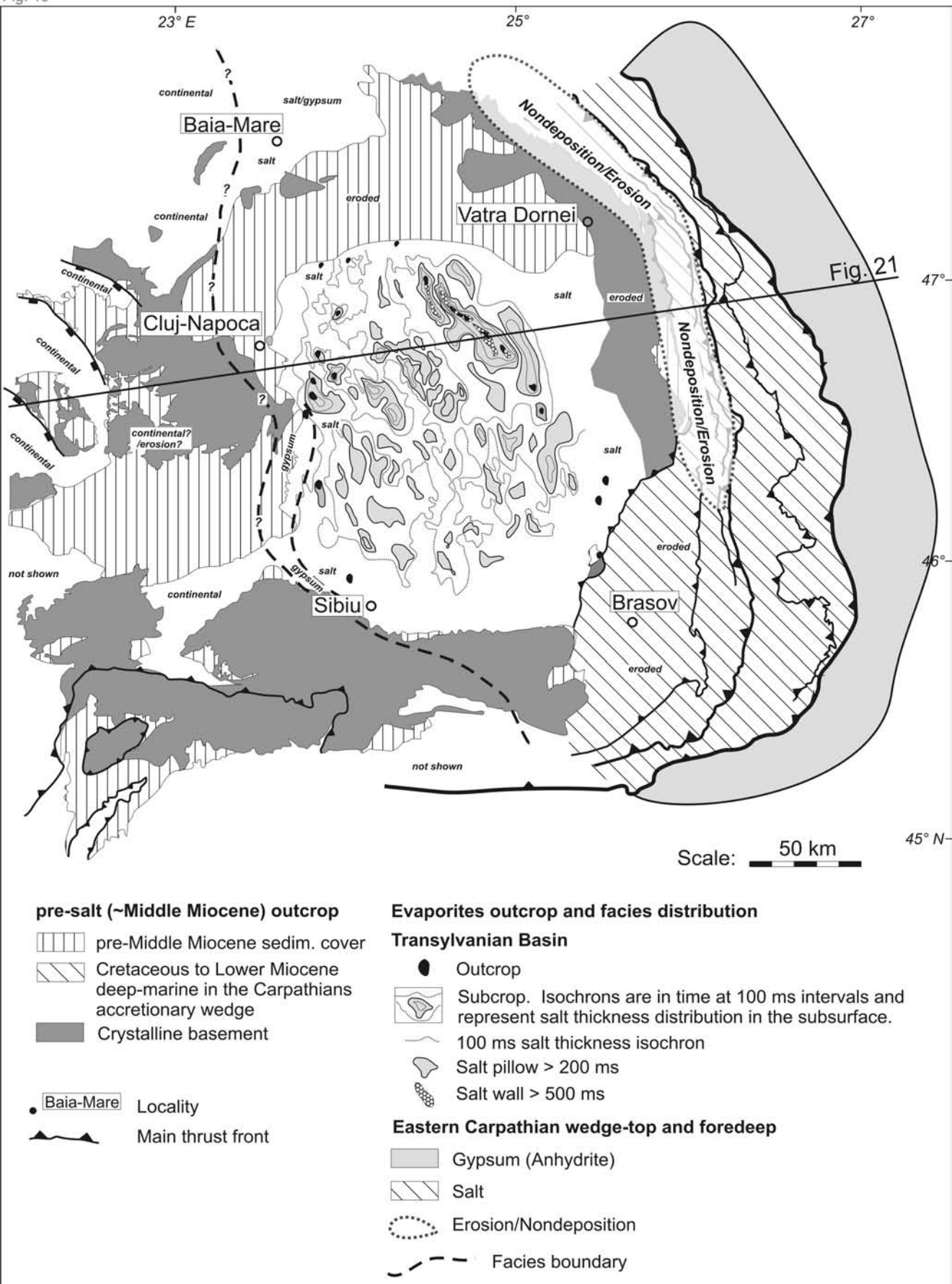
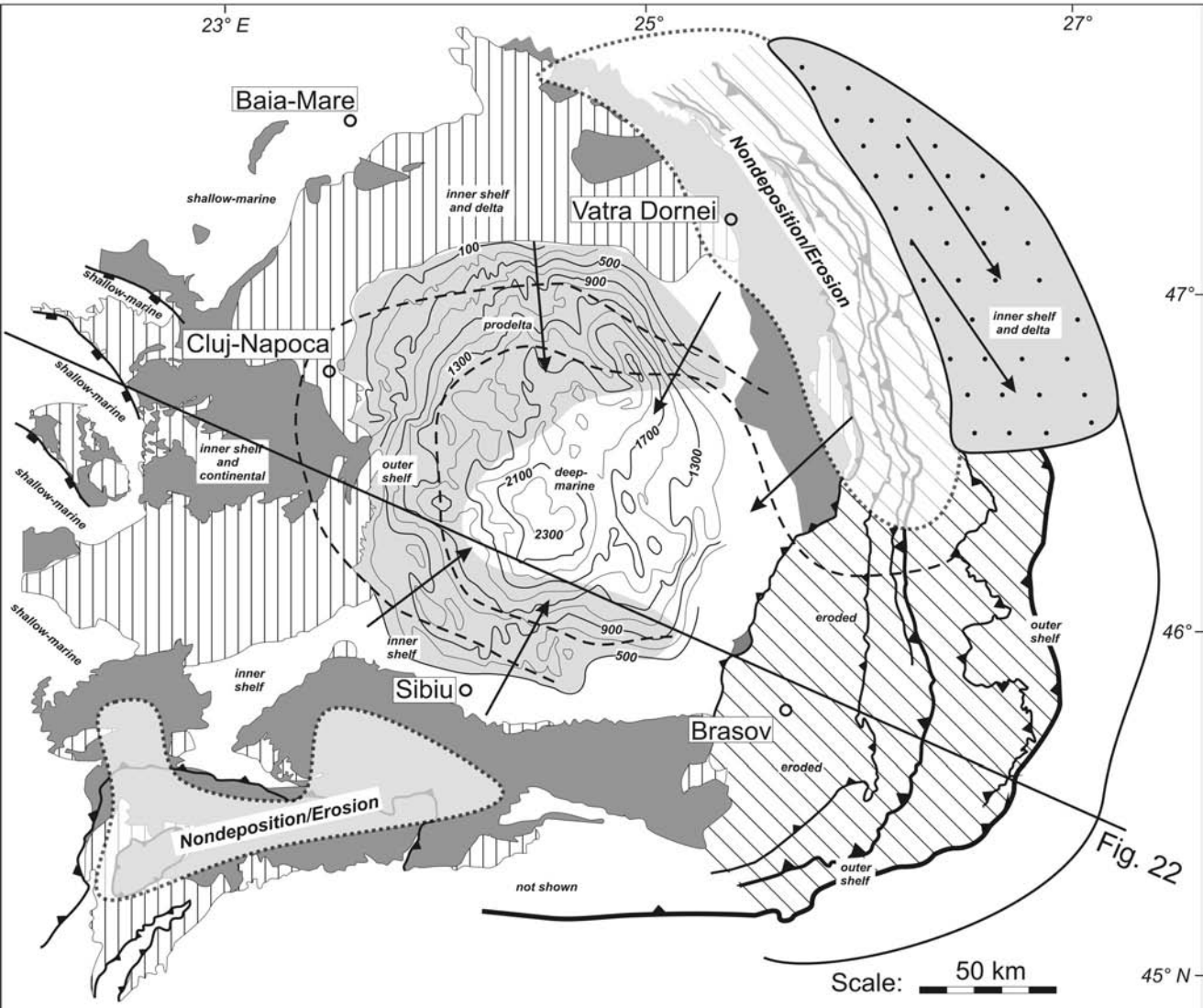

23° E
25°
27°
47°
46°
45° N
Baia-Mare
Vatra Dornei
Cluj-Napoca
Sibiu
Brasov
Fig. 21
continental
salt/gypsum
salt
eroded
continental?
/erosion?
gypsum
not shown
Nondeposition/Erosion
Scale:
50 km
pre-salt (~Middle Miocene) outcrop
pre-Middle Miocene sedim. cover
Cretaceous to Lower Miocene deep-marine in the Carpathians accretionary wedge
Crystalline basement
Baia-Mare
Locality
Main thrust front
Evaporites outcrop and facies distribution
Transylvanian Basin
Outcrop
Subcrop. Isochrons are in time at 100 ms intervals and represent salt thickness distribution in the subsurface.
100 ms salt thickness isochron
Salt pillow > 200 ms
Salt wall > 500 ms
Eastern Carpathian wedge-top and foredeep
Gypsum (Anhydrite)
Salt
Erosion/Nondeposition
Facies boundary

Fig. 20



Pre-salt (~Middle Miocene) outcrop

- pre-Middle Miocene sedim. cover
- Cretaceous to Lower Miocene deep-marine in the Carpathian accretionary wedge
- Crystalline basement

- Baia-Mare — Locality
- Main thrust front

Post-Salt thickness and facies distribution

- *inner/outer shelf, delta ...* — Middle Sarmatian facies distribution
- Middle Sarmatian paleoflow directions

Transylvanian Basin

- Post-salt thickness distribution with isochrons at 200 ms intervals. The isochrons are truncated by Pliocene erosion!
- Upper Badenian onlap distribution

Eastern Carpathian wedge-top and foredeep

- Inner shelf & delta
- Outer shelf/deep marine
- Erosion/Nondeposition
- Facies boundary

Fig. 21

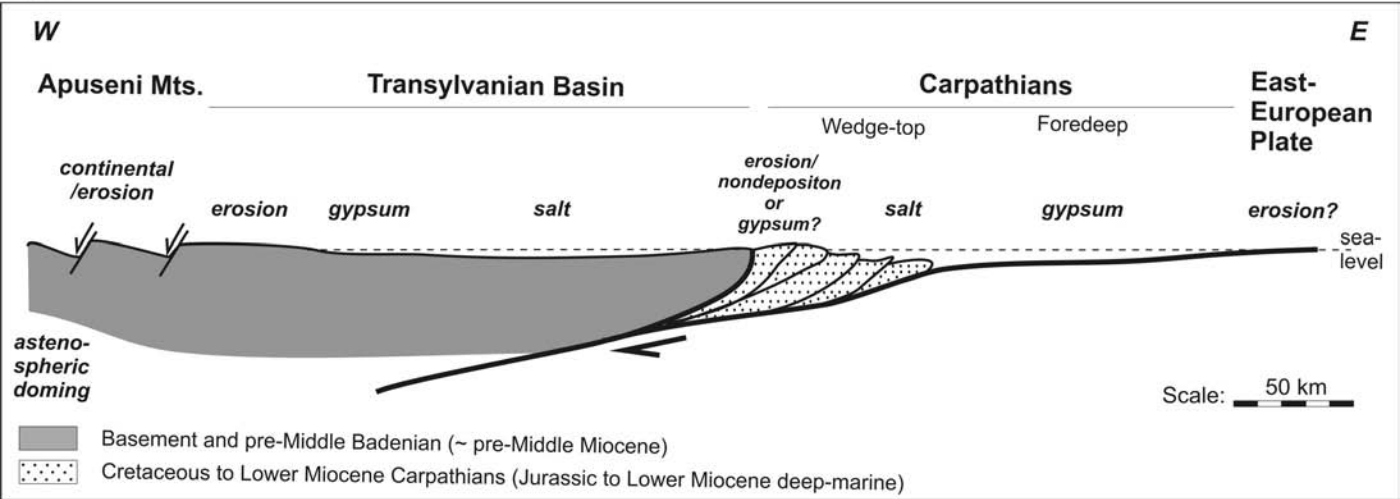

W
E
Apuseni Mts.
Transylvanian Basin
Carpathians
East-European Plate
Wedge-top
Foredeep
continental /erosion
erosion
gypsum
salt
erosion/ nondepositon or gypsum?
salt
gypsum
erosion?
sea-level
asteno-spheric doming
Scale: 50 km
Basement and pre-Middle Badenian (~ pre-Middle Miocene)
Cretaceous to Lower Miocene Carpathians (Jurassic to Lower Miocene deep-marine)

Fig. 22

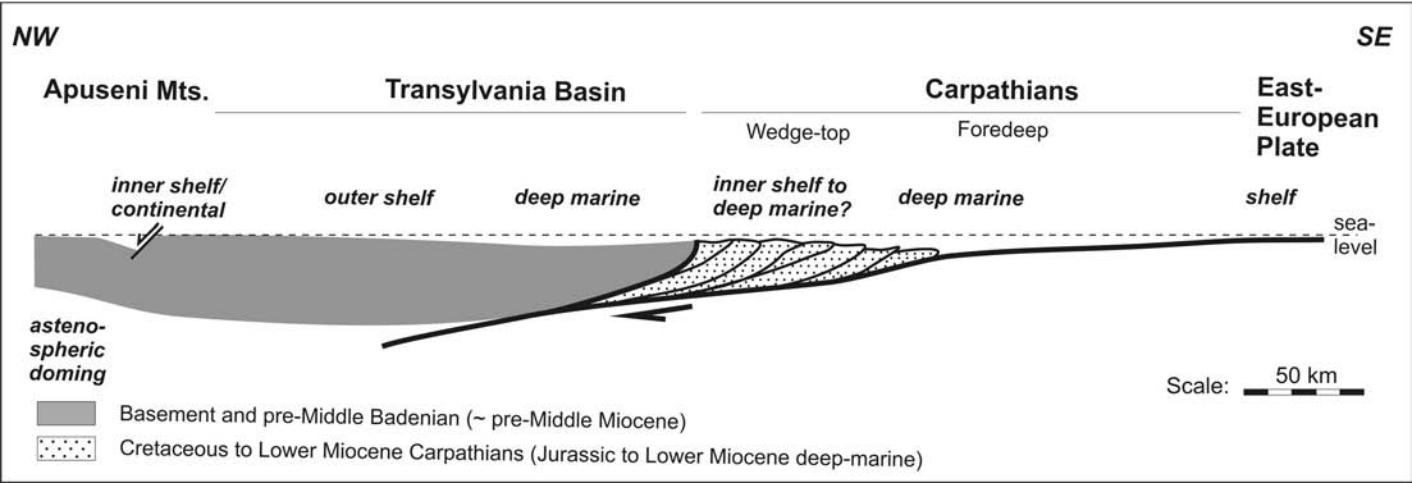

NW
SE
Apuseni Mts.
Transylvania Basin
Carpathians
East-European Plate
Wedge-top
Foredeep
inner shelf/ continental
outer shelf
deep marine
inner shelf to deep marine?
deep marine
shelf
sea-level
asteno-spheric doming
Scale:
50 km
Basement and pre-Middle Badenian (~ pre-Middle Miocene)
Cretaceous to Lower Miocene Carpathians (Jurassic to Lower Miocene deep-marine)

Fig. 23

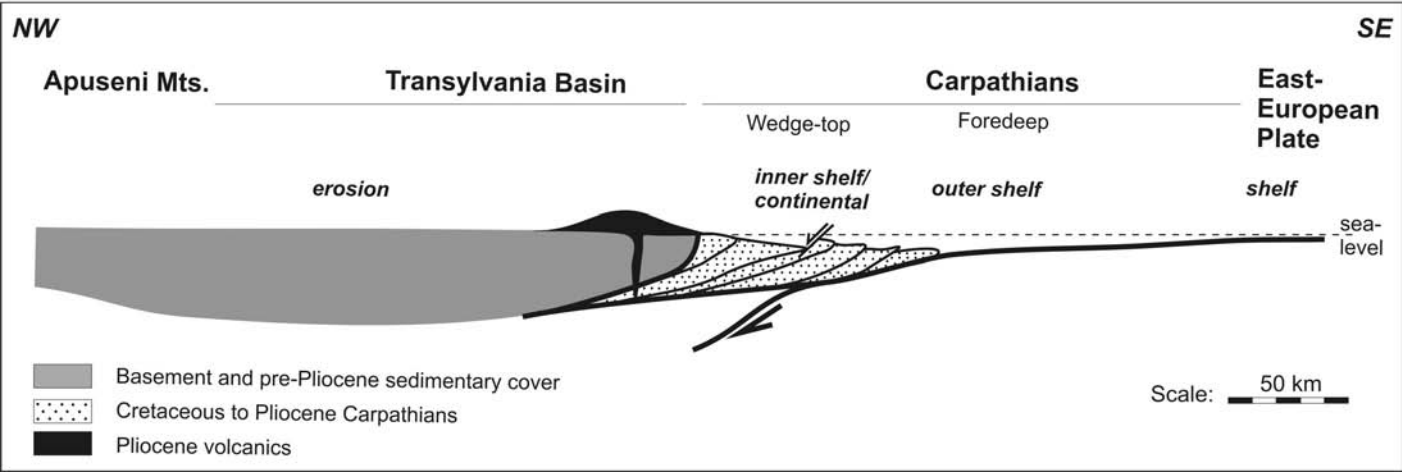
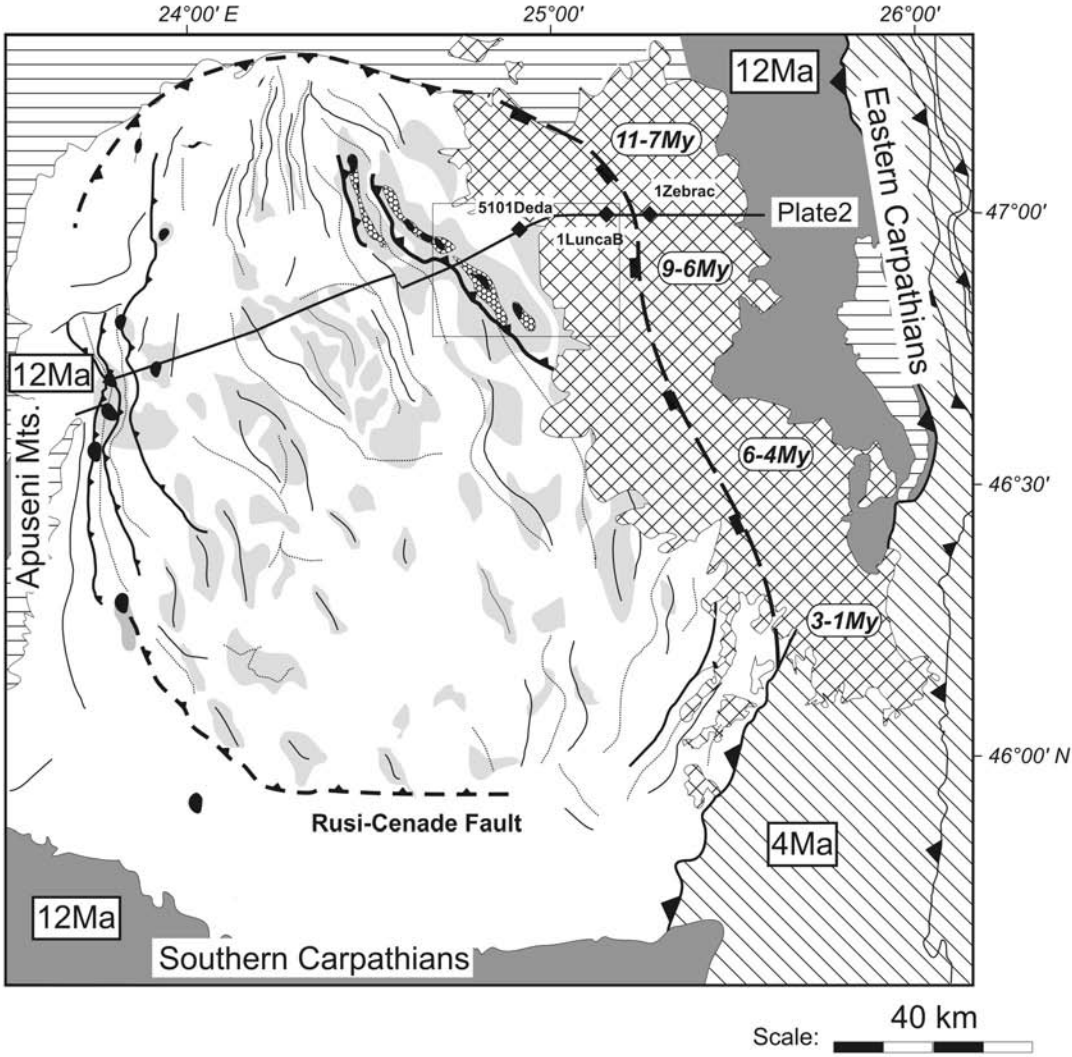

NW
SE
Apuseni Mts.
Transylvania Basin
Carpathians
East-European Plate
Wedge-top
Foredeep
erosion
inner shelf/ continental
outer shelf
shelf
sea-level
Basement and pre-Pliocene sedimentary cover
Cretaceous to Pliocene Carpathians
Pliocene volcanics
Scale:
50 km

Fig. 24



back-arc volcanics (post-Middle Miocene)

post-salt sedimentary cover (Upper Badenian to Pannonian)

pre-salt sedimentary cover (pre-Middle Badenian)

Carpathian accretionary wedge (post-Lower Cretaceous)

Crystalline basement

Carpathians

Major thrust front

12Ma Fission track uplift ages (Sanders, 1999)

11-7My Volcanism ages (Pécskay et al., 1995)

Transylvanian Basin

Syncline

Salt cored anticline

Main detachment scarp (listric fault)

Lateral ramp (thrust fault)

Toe thrust

Salt outcrop

Salt wall

Salt pillow thicker than 200 ms.

1Zebrac Well

Area shown by Fig. 25

Fig. 25

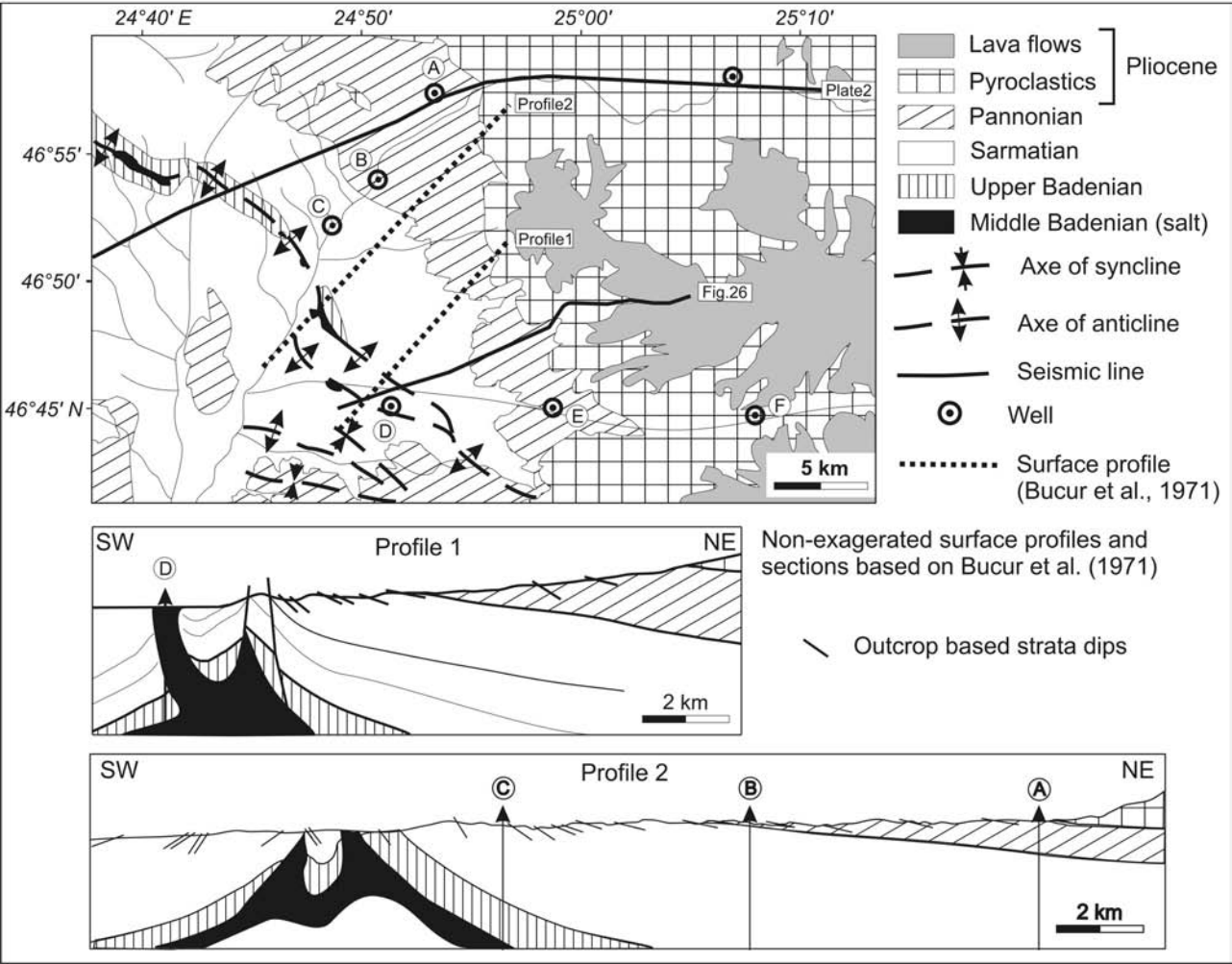

24°40' E
24°50'
25°00'
25°10'
46°55'
46°50'
46°45' N
A
B
C
D
E
F
Profile2
Profile1
Plate2
Fig.26
5 km
Lava flows
Pyroclastics
Pliocene
Pannonian
Sarmatian
Upper Badenian
Middle Badenian (salt)
Axe of syncline
Axe of anticline
Seismic line
Well
Surface profile (Bucur et al., 1971)
SW
Profile 1
NE
2 km
Non-exagerated surface profiles and sections based on Bucur et al. (1971)
Outcrop based strata dips
SW
Profile 2
NE
2 km

Fig. 26

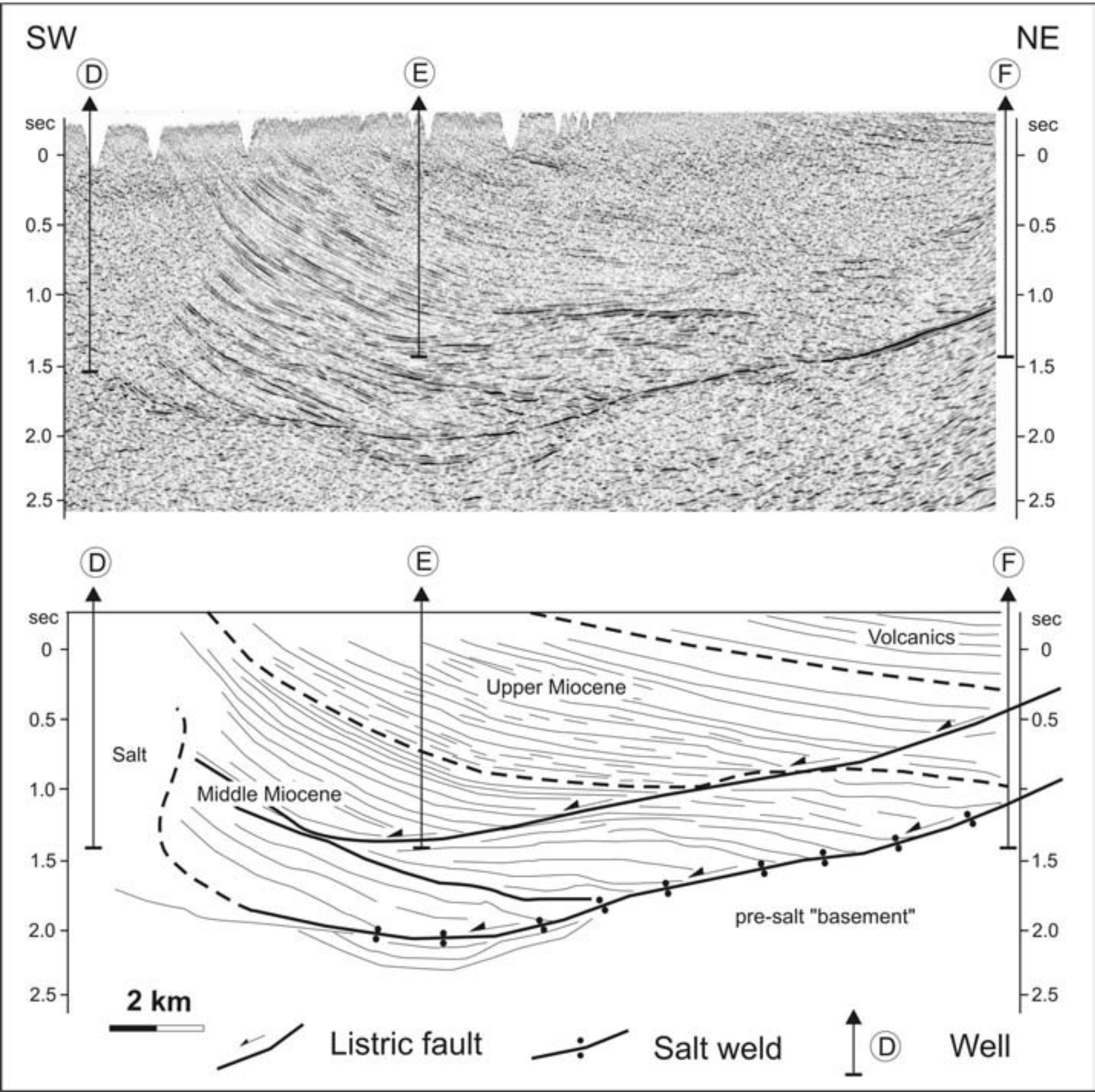

SW
NE
D
E
F
sec
0
0.5
1.0
1.5
2.0
2.5
Volcanics
Upper Miocene
Salt
Middle Miocene
pre-salt "basement"
2 km
Listric fault
Salt weld
Well

Fig. 27

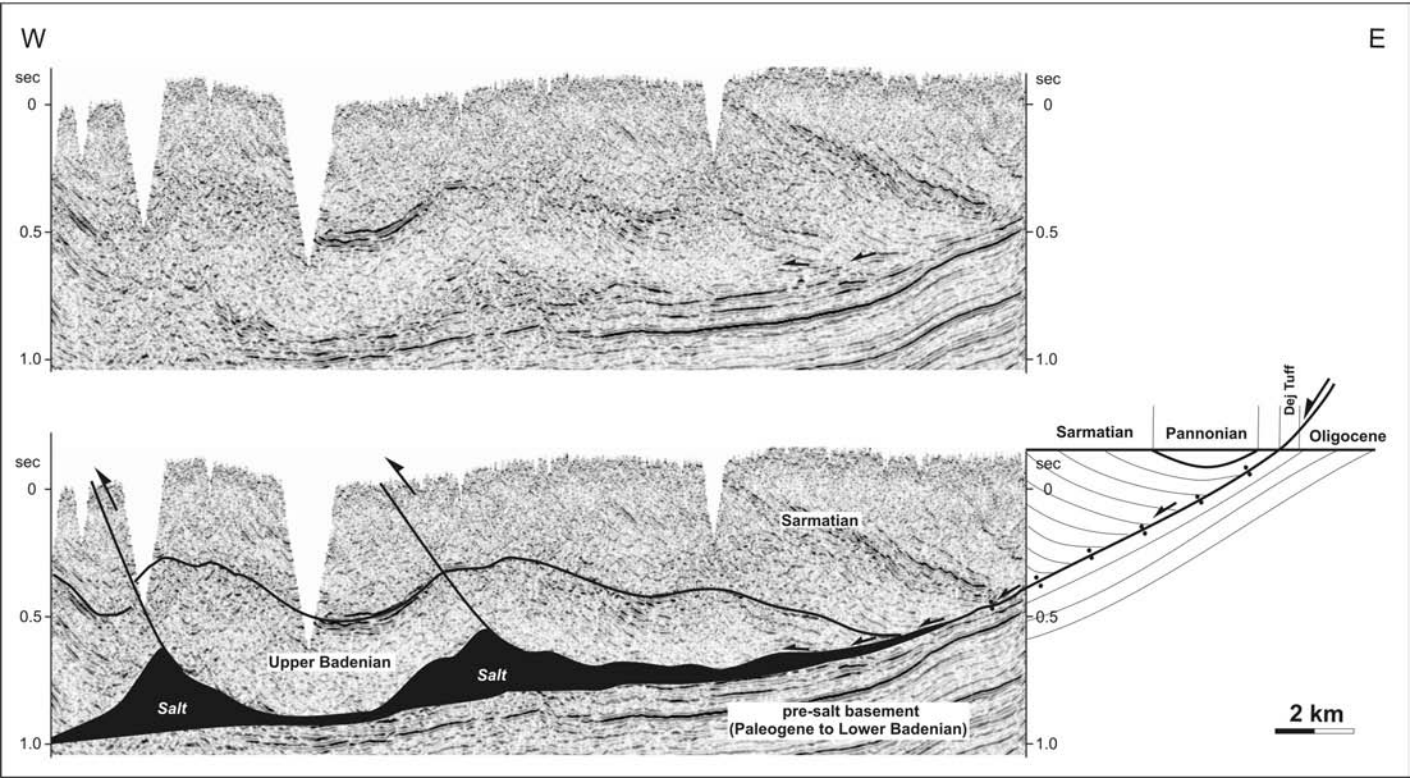
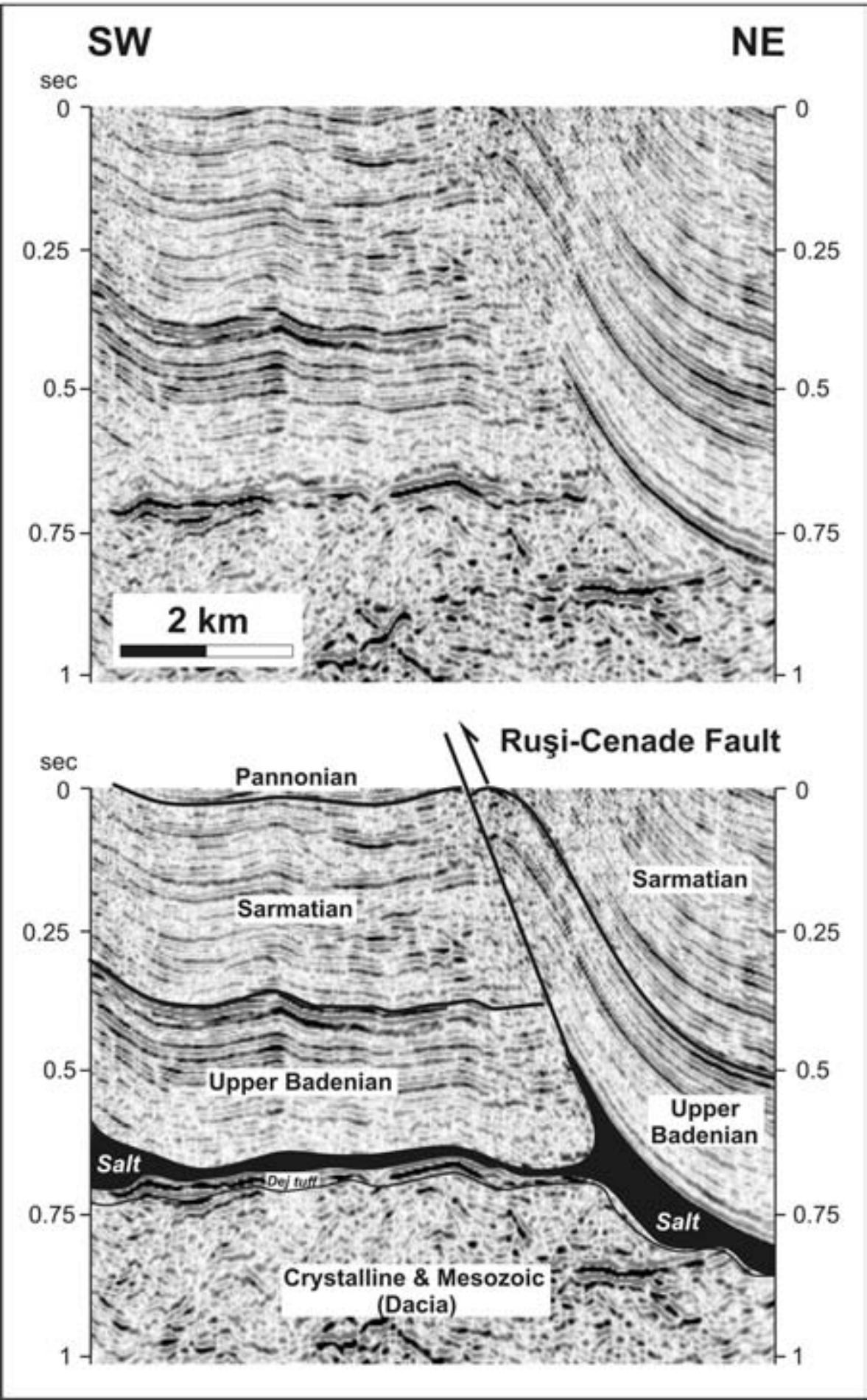
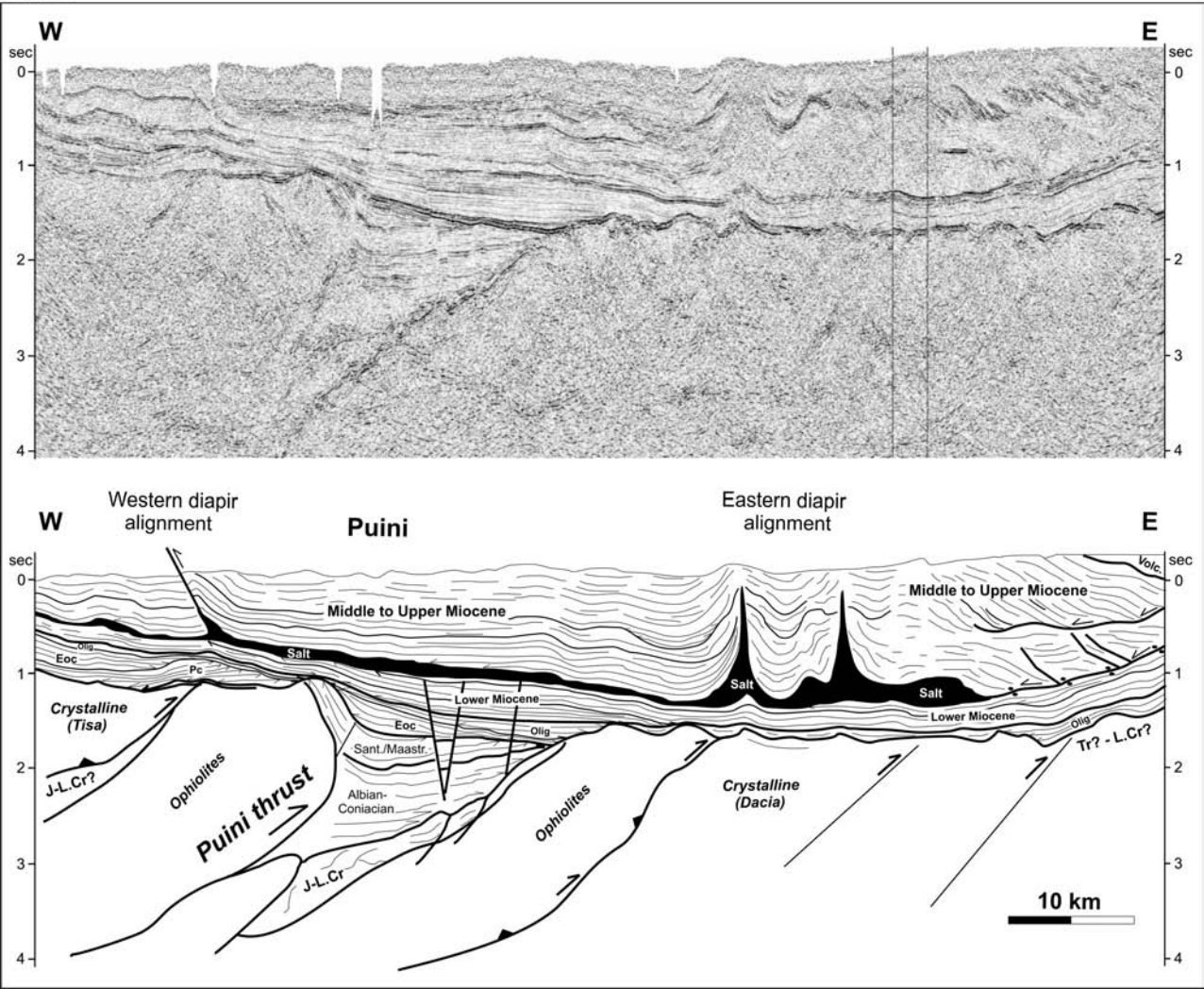
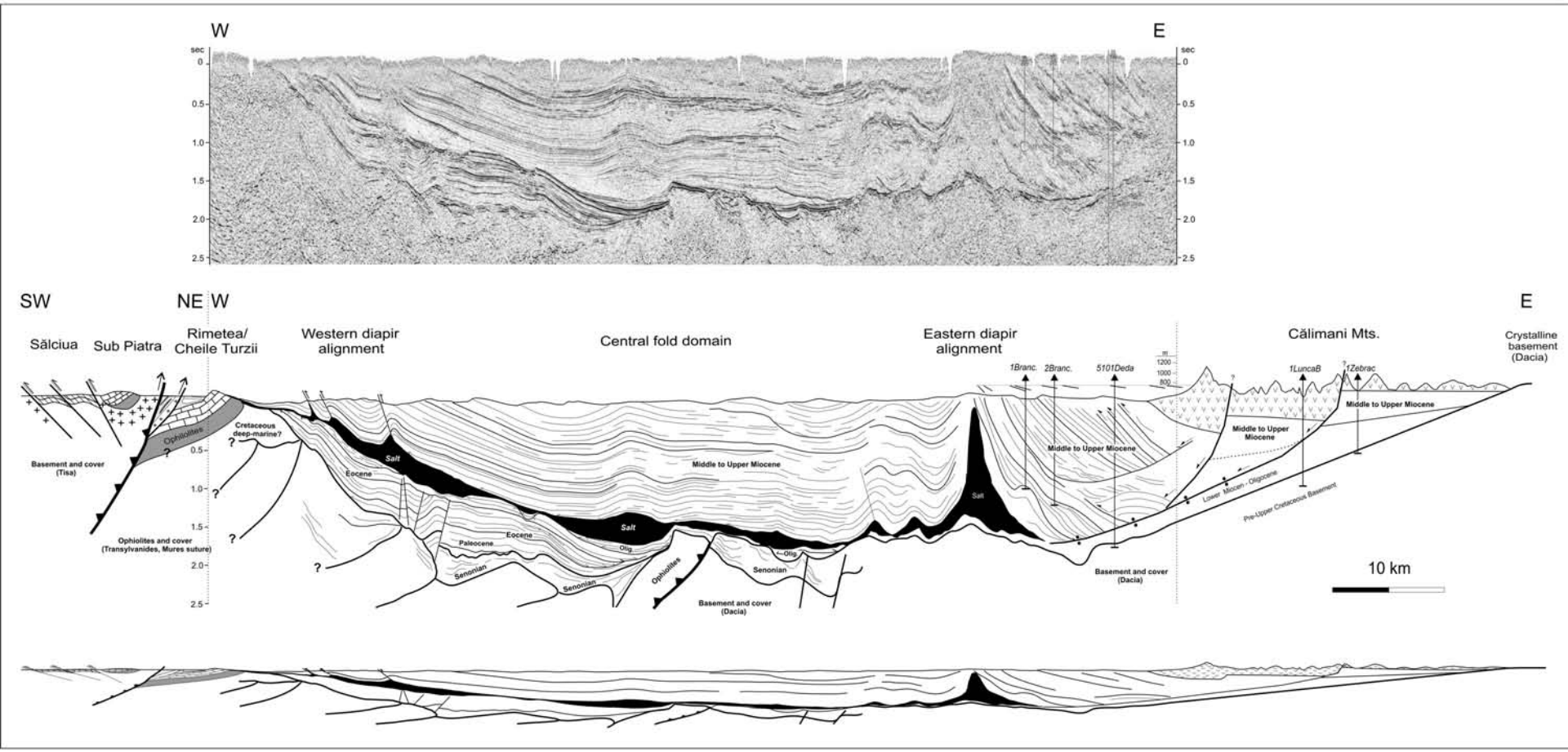

W
E
sec
0
0.5
1.0
Sarmatian
Upper Badenian
Salt
Salt
pre-salt basement
(Paleogene to Lower Badenian)
Sarmatian
Pannonian
Dej Tuff
Oligocene
2 km

Fig. 28


SW
NE
sec
0
0.25
0.5
0.75
1
2 km
Ruşi-Cenade Fault
Pannonian
Sarmatian
Sarmatian
Upper Badenian
Upper Badenian
Salt
Dej tuff
Salt
Crystalline & Mesozoic
(Dacia)


W
E
sec
0
1
2
3
4
Western diapir alignment
Puini
Eastern diapir alignment
Middle to Upper Miocene
Salt
Olig.
Eoc
Pc
Crystalline (Tisa)
J-L.Cr?
Ophiolites
Puini thrust
Lower Miocene
Eoc
Olig
Sant./Maastr.
Albian-Coniacian
J-L.Cr
Ophiolites
Crystalline (Dacia)
Salt
Salt
Lower Miocene
Volc.
Olig
Tr? - L.Cr?
10 km



Legend to Plates 1-6

Outcrop

- Ophiolites
- Crystalline
- Upper Jurassic and Lower Cretaceous limestones
- Albian-Aptian wildflysch
- Senonian
- Latest Miocene to Pliocene volcanics

Salt

Salt weld

4841ValeaL Well name

Middle Cretaceous major nappe fronts

Middle Cretaceous (tectonic transport to E)

Latest Cretaceous (tectonic transport to NW)

post-Late Miocene (Pliocene? or younger)

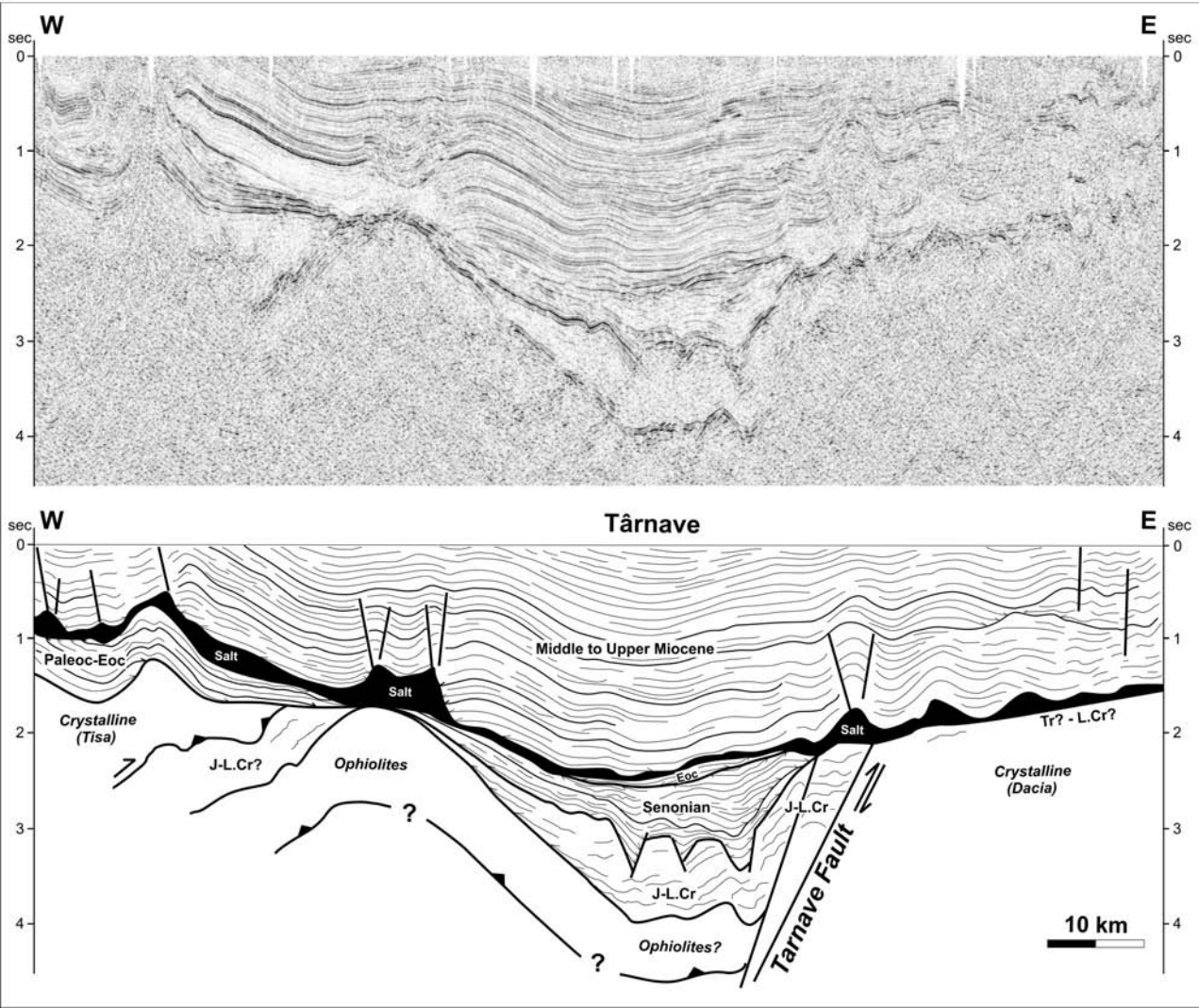
post-Late Miocene (Pliocene? or younger) thrust and listric faults related to gravitational spreading

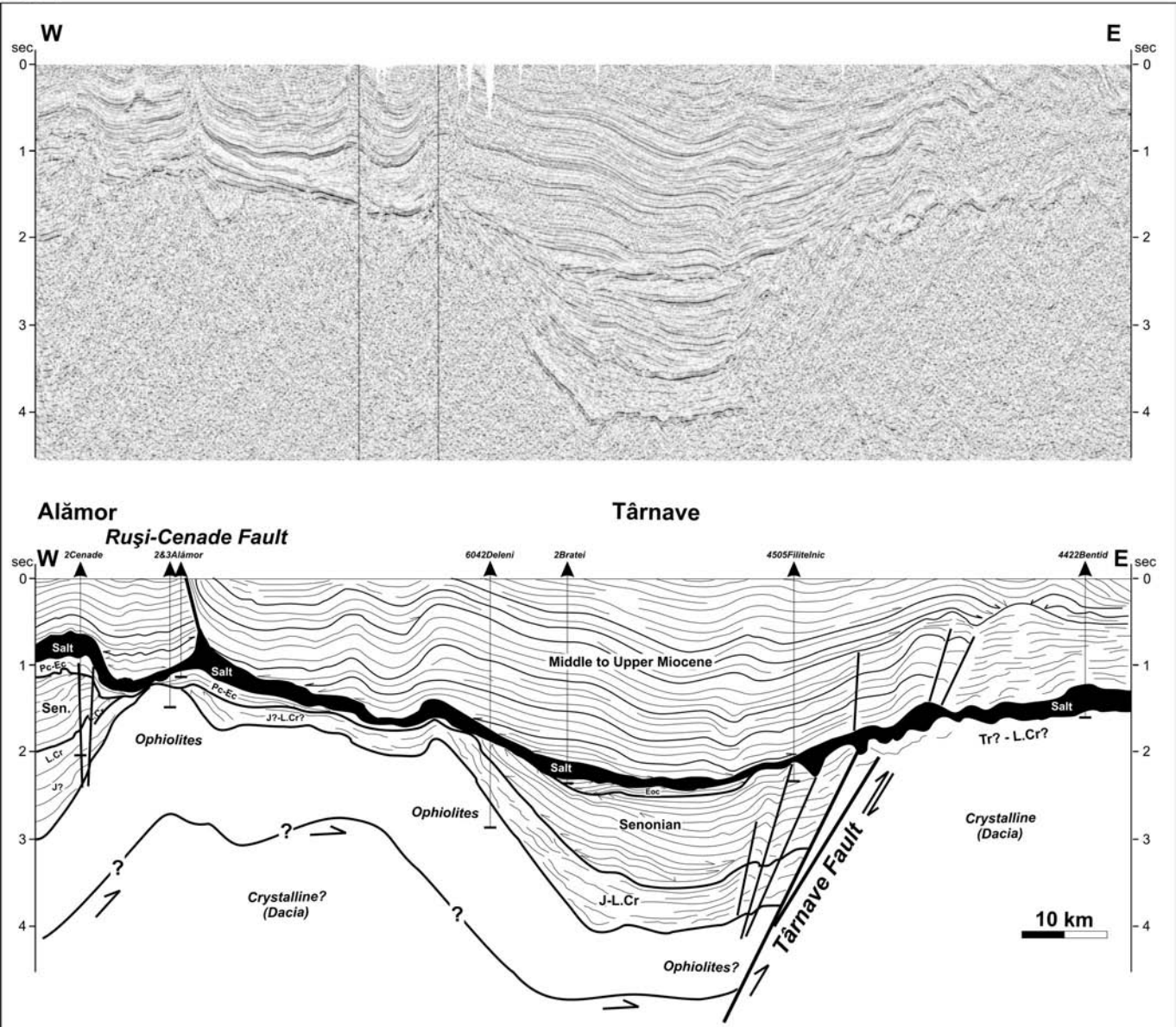
Projected outcrop based strata dips

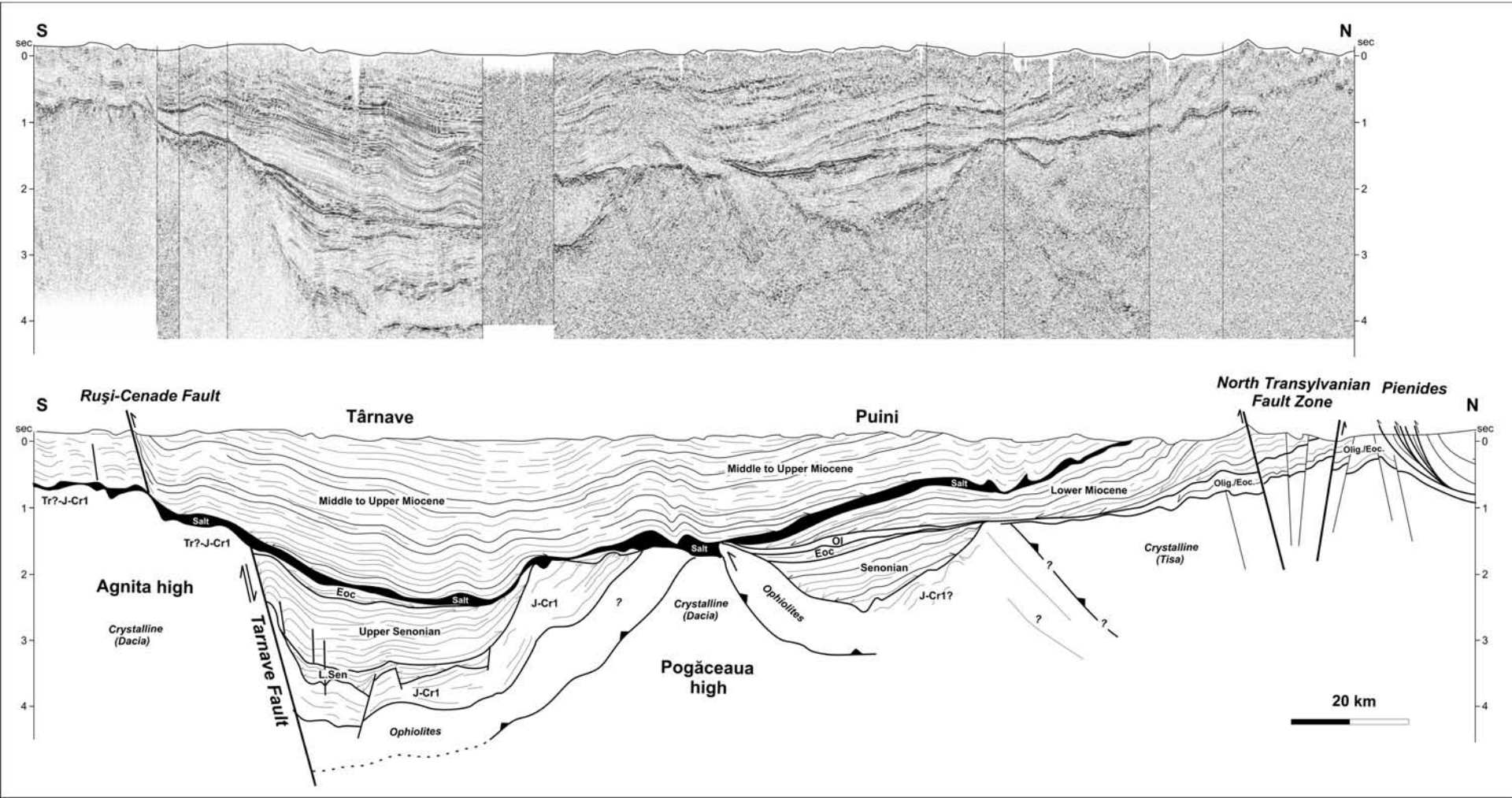
Upper Cretaceous normal fault inverted at the Cretaceous/Paleogene boundary

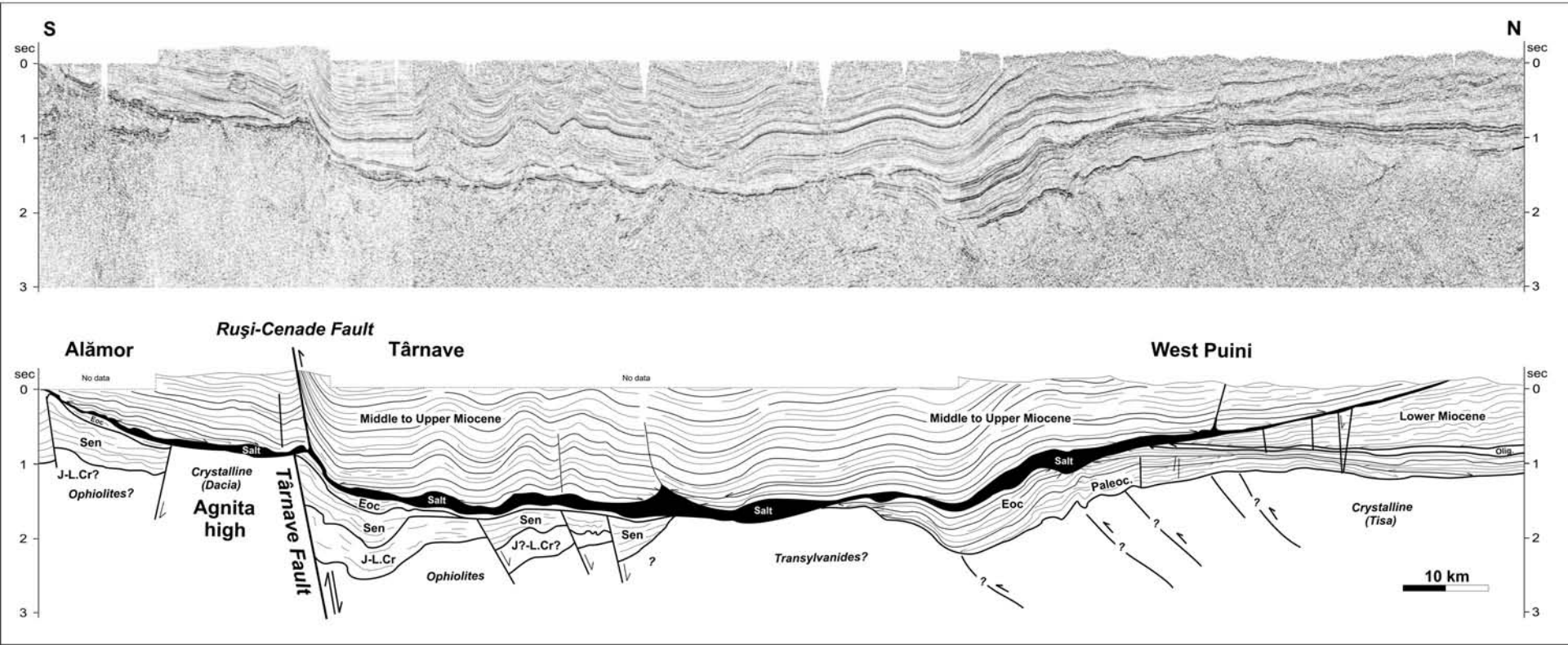
Abbreviations

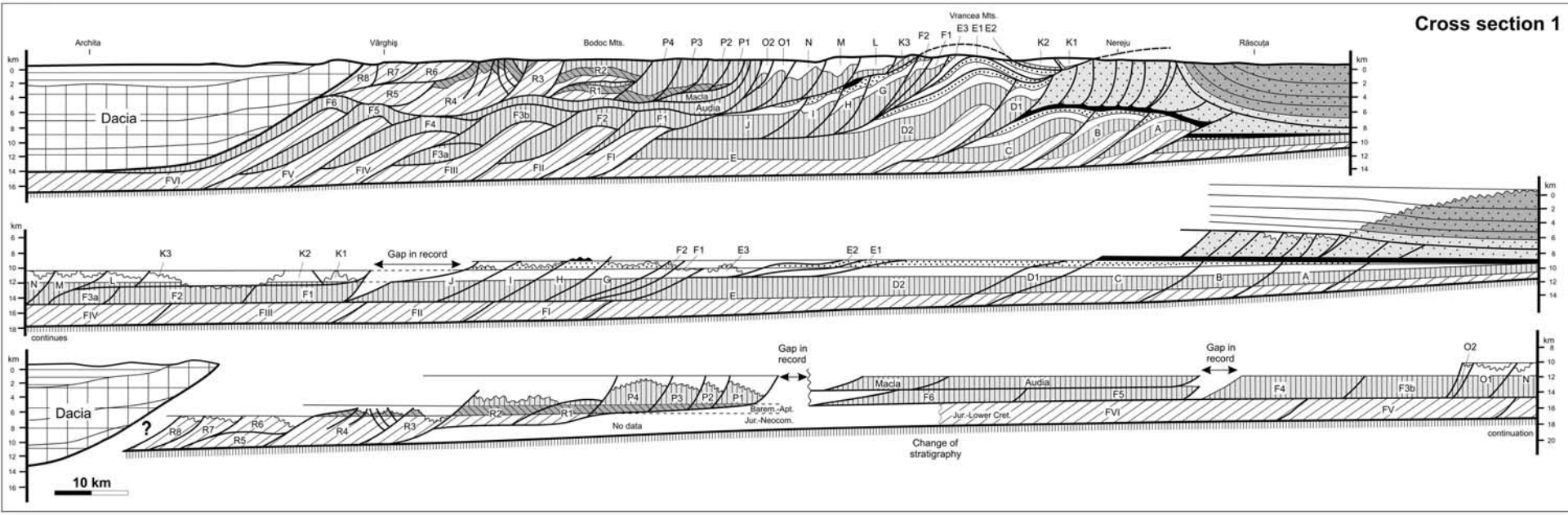
Olig.	Oligocene
Eoc.	Eocene
Paleoc.	Paleocene
Sen.	Senonian
Cr.	Cretaceous
J.	Jurassic
Tr.	Triassic
L.	Lower
U.	Upper


W
E
sec
0
1
2
3
4
Târnave
Paleoc-Eoc
Salt
Middle to Upper Miocene
Crystalline
(Tisa)
J-L.Cr?
Ophiolites
Eoc
Senonian
J-L.Cr
Tr? - L.Cr?
Crystalline
(Dacia)
Tarnave Fault
Ophiolites?
10 km


W
E
sec
0
1
2
3
4
Alămor
Târnave
Ruşi-Cenade Fault
2Cenade
2&3Alămor
6042Deleni
2Bratei
4505Filitelnic
4422Bentid
Salt
Pc-Ec
Sen.
L.Cr
J?
Ophiolites
J?-L.Cr?
Middle to Upper Miocene
Eoc
Senonian
J-L.Cr
Tr? - L.Cr?
Crystalline?
(Dacia)
Crystalline
(Dacia)
Ophiolites?
Târnave Fault
10 km


S
N
sec
0
1
2
3
4
Ruşi-Cenade Fault
Târnave
Puini
North Transylvanian Fault Zone
Pienides
Middle to Upper Miocene
Lower Miocene
Olig./Eoc.
Salt
Tr?-J-Cr1
Agnita high
Crystalline (Dacia)
Crystalline (Tisa)
Eoc
Ol
Senonian
J-Cr1
J-Cr1?
Upper Senonian
L.Sen
Tarnave Fault
Ophiolites
Pogăceaua high
20 km


S
N
sec
0
1
2
3
Ruşi-Cenade Fault
Alămor
Târnave
West Puini
No data
No data
Middle to Upper Miocene
Middle to Upper Miocene
Lower Miocene
Olig.
Eoc.
Sen
Salt
J-L.Cr?
Ophiolites?
Crystalline
(Dacia)
Agnita
high
Târnave Fault
Eoc.
Salt
Sen
J-L.Cr
Ophiolites
Sen
J?-L.Cr?
Sen
?
Salt
Transylvanides?
Salt
Eoc
Paleoc.
Crystalline
(Tisa)
10 km


Cross section 1
Archita
Vârghiş
Bodoc Mts.
Vrancea Mts.
Nereju
Răscuţa
Dacia
Macla
Audia
Gap in record
continues
Barem.-Apt.
Jur.-Neocom.
No data
Jur.-Lower Cret.
Change of stratigraphy
continuation
10 km

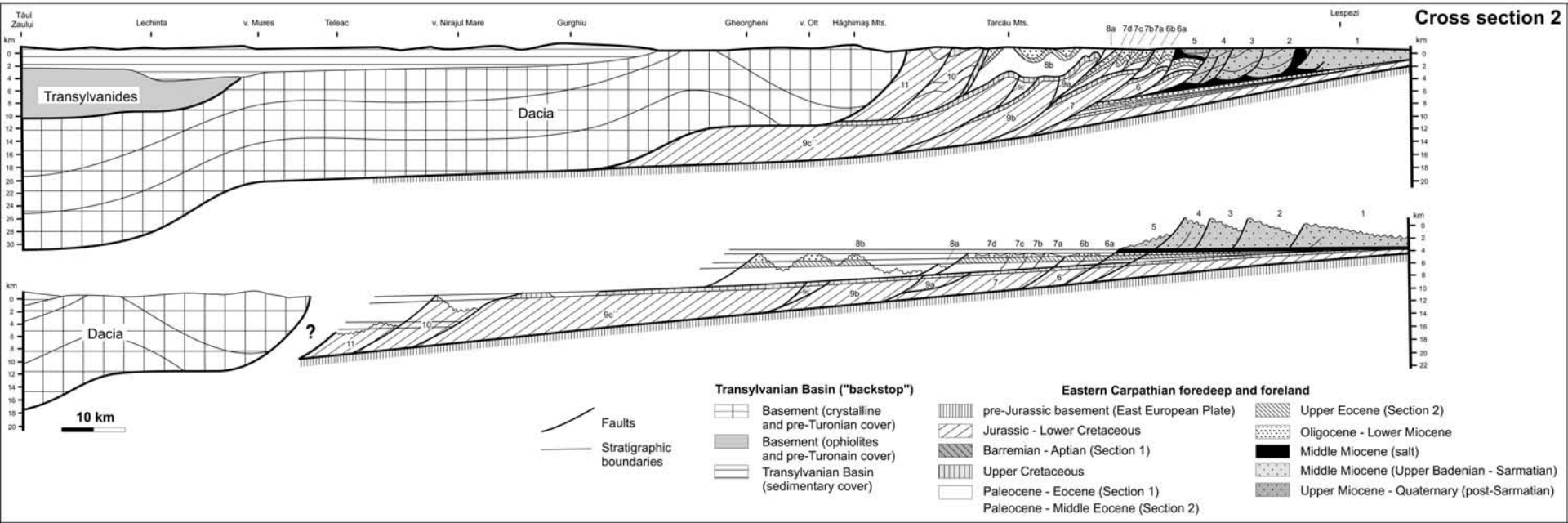

Cross section 2
Tăul Zaului
Lechinta
v. Mures
Teleac
v. Nirajul Mare
Gurghiu
Gheorgheni
v. Olt
Hăghimaş Mts.
Tarcău Mts.
Lespezi
Transylvanides
Dacia
10 km
Faults
Stratigraphic boundaries
Transylvanian Basin ("backstop")
Basement (crystalline and pre-Turonian cover)
Basement (ophiolites and pre-Turonain cover)
Transylvanian Basin (sedimentary cover)
Eastern Carpathian foredeep and foreland
pre-Jurassic basement (East European Plate)
Jurassic - Lower Cretaceous
Barremian - Aptian (Section 1)
Upper Cretaceous
Paleocene - Eocene (Section 1)
Paleocene - Middle Eocene (Section 2)
Upper Eocene (Section 2)
Oligocene - Lower Miocene
Middle Miocene (salt)
Middle Miocene (Upper Badenian - Sarmatian)
Upper Miocene - Quaternary (post-Sarmatian)

Table 1

No	Well name	Type of basement	No	Well name	Type of basement
1	1 Aiud	Late Cretaceous? granite	23	2 Săsăuşi	Crystalline
2	1 Cenade	Jurassic (Triassic?) basalt	24	2 Sic	Crystalline
3	1 Coaş Ineu	Crystalline	25	3 Lujerdiu	Crystalline
4	1 Dîrja	Crystalline	26	4 Bădeşti	Crystalline
5	1 Ghijasa	Triassic limestone	27	401 Zalău	Crystalline
6	1 Grînari	Lower Cretaceous, crystalline	28	413 Zalău	Crystalline
7	1 Jibert	Triassic limestone	29	4135 Bistrița	Crystalline
8	1 Lunca Bradului	Crystalline	30	4141 Beudiu	Jurassic basalt
9	1 Miheş de Campie	Jurassic basalt	31	4204 Ocna Mureş	Jurassic basalt
10	1 Nucet	Crystalline	32	4205 Viişoara	Jurassic? volcanics
11	1 Ocna de Sus	Crystalline	33	4401 Mercheaşa	Lower Cretaceous sandstone
12	1 Şalcău	Lower Cretaceous limestone	34	4411 Benţid	Crystalline
13	1 Săsăuşi	Crystalline	35	4501 Band	Jurassic limestone
14	1 Sic	Crystalline	36	457 Ileanda	Crystalline
15	1 Strîmbu	Crystalline	37	4901 Stînceni	Crystalline
16	1 Stupini	Crystalline	38	4912 Ibăneşti	Triassic limestone
17	1 Zetea	Crystalline	39	4913 Ibăneşti	Permian(?) conglomerate
18	1 Zoreni	Jurassic basalt	40	5 Pogăceaua	Crystalline
19	10 Ilimbav	Crystalline	41	5101 Deda	Crystalline
20	1041 Gurghiu	Crystalline	42	6042 Deleni	Jurassic basalt
21	2 Alămor	Lower Cretaceous sandstone	43	7 Agnita	Triassic limestone
22	2 Ilimbav	Crystalline	44	9 Ucea	Triassic limestone