Volcano–basement interaction in the Eastern Carpathians: Explaining unusual tectonic features in the Eastern Transylvanian Basin, Romania

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Abstract

Evidence of volcano–basement interaction related to the Neogene/Quaternary Călimani-Gurghiu-Harghita volcanic chain (CGH) in the Eastern Carpathians, Romania, includes both structural and morphological features. The CGH is located roughly between the Carpathian thrust-and-fold arc to the east and the Transylvanian Basin to the west. Geological and geophysical data demonstrate basement subsidence along most of the axial part of the volcanic chain. The wide volcaniclastic plateau developed preferentially westward by adjoining composite volcanoes, extends deep into the Transylvanian Basin, and its surface is tilted toward the chain axis, especially in the Gurghiu Mountains. These features strongly suggest late-stage or post-volcanic basement sagging beneath the volcanoes and related peripheral uplift. Interpretation of 2D seismic profiles acquired in the Transylvanian Basin indicates strong influence of the salt-related tectonics on the entire post-salt sequence. The Middle to Late Miocene succession, including the volcanic edifices, are tilted toward the basin margins. The pre-salt deposits together with the metamorphic basement are not tilted; instead, they are progressively uplifted. We suggest that the salt-related tectonics enhanced by volcano-induced compressive tectonics is the main triggering factor. The Neogene salt-tectonics developed under an overall regional compressive stress field. The salt withdrawal beneath the Upper-Miocene siliciclastites was amplified by the influence of the weight of the volcanic edifices. The post-salt succession (including the volcanics) tilting is related to salt withdrawal processes and rotation of the whole post-salt sedimentary succession along the salt-layer acting as a detachment surface. Swarms of small-scale west-verging reverse faults developed near the CGH volcanic range represent small-scale thrust structures accommodating the rotation induced compression in the post-salt brittle lithologies. The observed tectonic features are characteristic only for the eastern part of the Transylvanian Basin. No such features have been identified along the western margin or elsewhere in the basin. The peripheral deformation of the pre-volcanic shallow sedimentary basement induced sagging and spreading of the nearby large volcanic edifices. This process combined with the effect of increased heat-flux may have strongly enhanced salt diapirism as well. Volcano spreading was asymmetric because the basement at the eastern margin of the volcanic chain consists of brittle Mesozoic sedimentary and Precambrian metamorphic rocks of the Eastern Carpathian thrust-and-fold belt. The buttressing effect to the east favored spreading toward the west. Tectonic features resulted from volcano–basement interaction are combined with those related to regional tectonic processes involved in the evolution of the Transylvanian Basin and the Eastern Carpathians. Interplay between

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regional stress fields and local volcano-induced stress may best account for the complexity of tectonic features recorded in the eastern Transylvanian Basin.

1. Introduction

The recognition of dynamic interaction between volcanoes and their basement is a recent development in volcanology (e.g., Delaney, 1992; Borgia, 1994; Merle and Borgia, 1996). Basement deformation under the weight of large volcanoes in turn induces deformation within the edifices themselves which influences further eruptive behavior.

The main way the edifice adjusts to deformation of weak basement underneath is volcano spreading, involving subsidence of the central summit part and centrifugal lateral expansion at its peripheries as it was demonstrated via scaled experiments (Merle and Borgia, 1996). To accommodate space reduction due to sagging, plastically deformable basement rocks are squeezed from beneath the central part of volcanoes and displaced laterally, while the resulting compressive stress is resolved through reverse faulting and thrusting at the distal peripheries. It is obvious that volcano–basement interaction is strongly dependent on basement rock types and structure.

Volcano spreading as resulting from interaction between edifice and basement has been demonstrated at a number of active or recent composite volcanoes worldwide (e.g., Borgia et al., 1990; Van Wyk De Vries and Borgia, 1996; Merle and Borgia, 1996). However, relevant examples of volcano–basement interaction are as yet missing for older, partially eroded volcanoes, where volcano-induced tectonic structures can more readily be investigated and understood. This paper discusses the particular case of interaction between composite volcanoes of the Neogene Eastern Carpathian volcanic range and their Miocene sedimentary basement making of the fill of the Transylvanian Basin, Romania, where deformation of both volcanic edifices and basement is recorded. This process may account for puzzling tectonic features recognized in the eastern part of the Transylvanian Basin, which do not fit with basin evolution models proposed to date.

2. Geological setting

The Transylvanian Basin and the Eastern Carpathians are both parts of the larger alpine Car-pathian–Pannonian orogenic system (Fig. 1), which developed at the western margin of the Eurasian Plate involving at least two smaller lithospheric blocks, the Tisia–Dacia and the ALCAPA blocks (Balla, 1987; Csontos et al., 1992; Csontos, 1995). Eastward translation (Ratschbacher et al., 1991; Csontos et al., 1992; Fodor et al., 1999) and divergent rotation (Pătrașcu et al., 1994; Márton and Fodor, 1995) of these blocks resulted in the arcuate shape of the Carpathians (Mațenco, 1997; Zweigel, 1997). The main shortening phases occurred during Middle Creta-ceous and Miocene times (Săndulescu, 1988) accom-panying subduction of the Tisia–Dacia and ALCAPA blocks beneath the Eurasian Plate (Balla, 1987; Csontos et al., 1992).

Intense and widespread magmatic activity accompa-nied deformation in the area during Miocene to Upper Pleistocene (Pécskay et al., 1995). Both calc-alkaline and alkali-basaltic compositions are present. Most of the products of the mostly andesitic calc-alkaline magma-tism occur as a subduction-related volcanic arc, typically located along the inner part of the arcuate Carpathian fold-and-thrust belt.

The Transylvanian Basin develops between the East Carpathians, South Carpathians and the Apuseni Mountains (Fig. 1) in a back-arc type tectonic setting (Burchfiel and Royden, 1982; Balintoni et al., 1997; Sanders, 1999; Huismans, 1999; Ciulavu, 1999) with respect the Carpathian arc (Săndulescu, 1988; Mațenco, 1997; Gârbacea, 1997; Zweigel, 1997; Sanders, 1999).

The sedimentary succession comprises Upper Me-sozoic to Miocene sediments deposited in several subsequent basin types (Balintoni et al., 1998). The Middle to Upper Miocene sediments are mainly represented by siliciclastics (see Fig. 5). One of the characteristic features of the basin fill is the Middle Miocene evaporites, represented by shallow-water gypsum (Ghergari et al., 1991) and deeper-marine salt (Dragoș, 1969; Stoica and Gherasie, 1981). The Middle to Upper Miocene sedimentary succession belongs to the latest megasequence (2nd order sedimentary cycle) of the basin fill and can be divided into eight 3rd order depositional sequences (Krézsek and Filipescu, 2005).

3. The Călimani-Gurghiu-Harghita volcanic chain

The Călimani-Gurghiu-Harghita volcanic chain (CGH hereafter), occurring along the eastern margin of the Transylvanian Basin (Fig. 1), is the southeastern segment of the Carpathian volcanic arc (Seghedi et al., 2004). It consists of a NW–SE-trending row of closely-spaced adjacent volcanic edifices. Most of them are medium-size composite volcanoes, a couple of them with summit calderas (Szakács and Seghedi, 1995). Obvious along-arc migration of volcanism from NW to SE has been pointed out (Rădulescu et al., 1972; Pécskay et al., 1995), accompanied by decreasing height, volume and complexity in the same direction (Szakács et al., 1993; Szakács and Seghedi, 1995).

Volcanic edifices in the CGH consist of lava-dominated central cones flanked by extensive volcaniclastic aprons. The central-type composite edifices were mostly built by more mafic (basalt and basaltic andesite)

effusive products including cone-forming lava flows and more viscous and felsic (andesite and dacite) lava domes concentrated in the summit areas of the volcanoes. Since the volcanoes are closely-spaced, their volcaniclastic aprons mostly developed sideways, especially westwards, towards the Transylvanian Basin where a flatland allowed unrestricted accommodation and redistribution of volcanic material from the central and proximal parts of the edifices. Mechanisms of primary and secondary dispersion of fragmented volcanic material of the volcaniclastic aprons included pyroclastic flows, debris flows, debris avalanches and fluvial erosion-transport systems. A widespread, gently outward-dipping volcanic ring-plain resulted on top of the latest (Pannonian) sedimentary fill of the eastern Transylvanian Basin. In contrast, volcaniclastic deposits show limited eastward extension due to the buttressing effect of the nearby Carpathian thrust-and-fold belt with a more rugged topography. Westward-dipping pre-
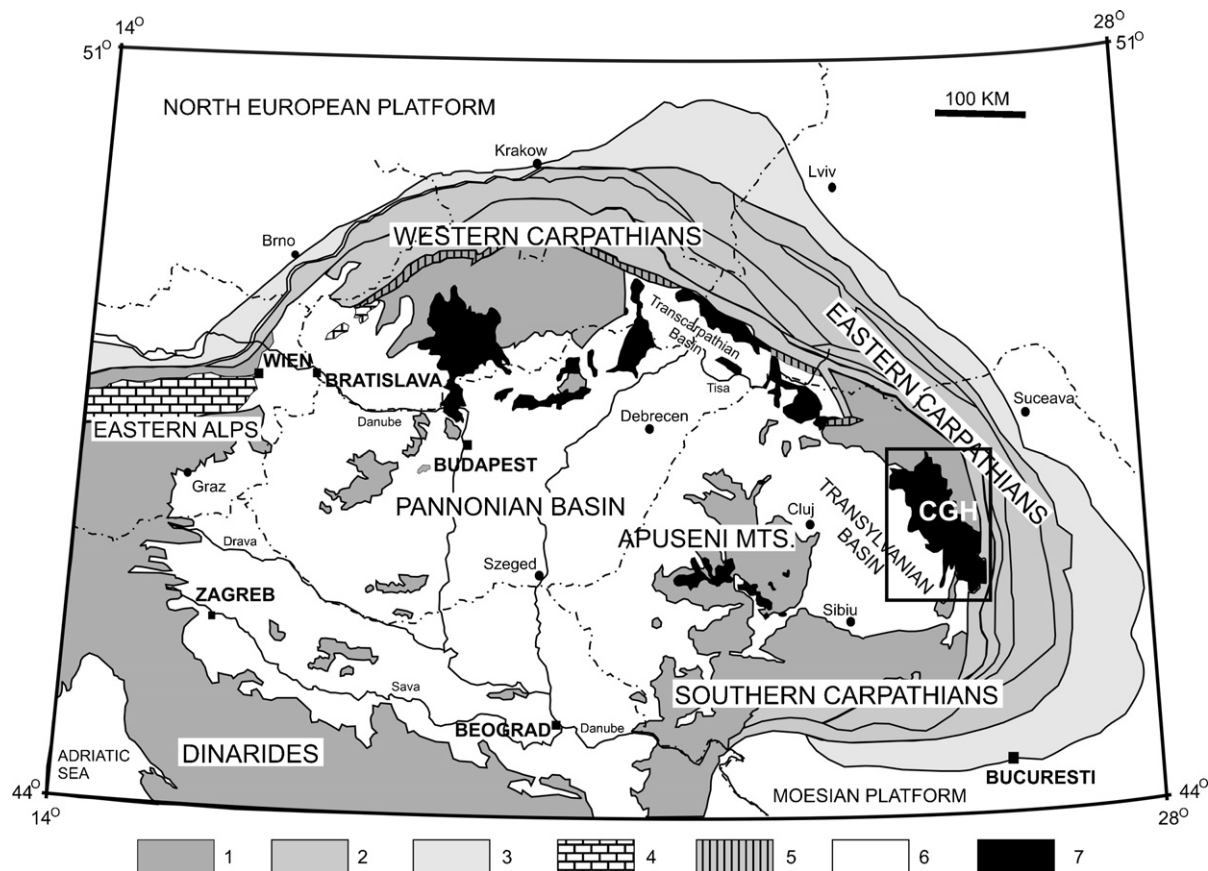

Fig. 1. Sketch map of the Carpathian–Pannonian Region (after Pécskay et al., 1995, modified) with location of the study area. 1 — Inner Carpathian units, 2 — Outer Carpathian Units, 3 — Carpathian Foredeep, 4 — Eastern Alps, 5 — Klippen Belt, 6 — Intra-Carpathian basins, 7 — Arc-type Neogene/Quaternary volcanics. Frame shows location of Fig. 2 with the Călimani-Gurghiu-Harghita volcanic chain (CGH).

volcanic topography also contributed to the asymmetrical development of volcaniclastic aprons along the CGH chain.

4. Volcano–basement interaction

Interaction between volcanic edifices and their pre-volcanic basement is recorded in a number of topographic and tectonic features that can be observed or inferred both inside and outside the CGH volcanic chain. Effects of such interaction on the volcanic edifices are obvious from geological and geophysical data and they strongly suggest volcano spreading. On the other hand, some unusual tectonic features, particular to the eastern margin of the Transylvanian Basin, may be interpreted as structures formed by a complex interplay of salt-tectonics, regional tectonics, and volcano-induced tectonics.

4.1. Volcano spreading in the CGH

Evidence of volcano spreading in the CGH is twofold: topographic and geologic.

4.1.1. Topographic evidence

Volcaniclastic ringplains around composite volcanoes show a gently outward-dipping overall topography as the smooth extension of the hyperbolic topographic profile of the volcanic edifices themselves. The morphological transition between volcaniclastics and basement is normally smooth or marked by a gentle slope-break.

One long-standing puzzling topographic feature of the CGH chain is the inward-dipping topographic and structural surface of the volcaniclastic apron developed at the western peripheries of the volcanoes of the Gurghiu Mts and of the northernmost part of the North Harghita Mountains. The highest topographic points of these surfaces are located along the westernmost margin of the volcaniclastic plateau, forming local peaks such as Bekecs, Siklód, Küsmöd, Firtos, Szarkakő (from north to south) (Fig. 3). From these peripheral high points the topographic surfaces show gentle to moderate dip northeastwards, i.e., towards the axial part of the CGH chain. A series of transversal topographic profiles (Figs. 2 and 3) and panoramic pictures (Fig. 4) clearly record this feature. The surfaces flatten and eventually reverse to the "normal" outward dip towards the limit between volcaniclastic medial facies and lava-dominated proximal facies (Fig. 3).

Inward-tilted surfaces of peripheral volcaniclastic formations can readily be observed along the southern 2/3 of the western Gurghiu Mountains between the Gurghiu and Târnava Mare valleys, where medial to distal volcanic facies is best developed (Figs. 2 and 3). North of the Gurghiu valley, erosional dissection is much more advanced because of tectonic uplift of the pre-Neogen basement along a seismically active fault (Ciulavu, 1999), hence structural surfaces are less well preserved and more difficult to reconstruct. South of the Târnava Mare valley the volcaniclastic plateau is generally less developed, and the volcanic material has been redistributed along tectonically determined south–southwest-trending paleovalleys parallel to the Perşani Mountains Neogene horst-and-graben structure. Except for its northwesternmost part, no well-individualized plateau-type topography developed peripherally adjacent to most of North Harghita and South Harghita Mountains (Fig. 2). Thus, no anomalous topographic features can be observed here, nor in the eastern side of the North Harghita Mountains, where extensive volcaniclastic plateau does also occur in places. The surfaces of peripheral volcaniclastic deposits here show normal outward-decreasing topographic gradients.

Inward-dipping structural surface of the peripheral volcaniclastic plateau can only be explained by inversion of the dip direction from outward to inward after most volcanic edifice construction was accomplished. Mechanisms which may be envisaged are tectonic, volcanic or both. Regional tectonic tilting of paleosurfaces may result from differential uplift and/or subsidence. Uplift of the westernmost peripheries of the volcaniclastic plateau is unlikely to be related to non-volcanic regional tectonic events, because no bounding faults are recognized nearby. Faults disrupting the paleosurface are located much inwards (Fig. 3). They bound segments of inward-dipping surfaces. Subsidence beneath the lowest-positioned paleosurface areas, on the other hand, can certainly be ruled out because beneath at least part of such areas (e.g., Sovata-Praid-Corund area) just the opposite effect can be expected due to Middle-Miocene to recent salt diapirism (Ciupagea et al., 1970; Visarion et al., 1976) (Fig. 3). Volcano spreading may conveniently account for inward-dipping of the volcaniclastic topographic surface. Sagging of the central parts of volcanic edifices in the Gurghiu Mountains and accompanying peripheral shortening by small-scale thrust-faulting is a likely mechanism for distal uplift of the volcanic plateau together with its sedimentary basement.

Andesite-dominated composite volcanoes, such as those in the CGH chain, are characterized by their cone-shaped morphology. Modern composite volcanic cones display upward-concave outer slopes with slope angles
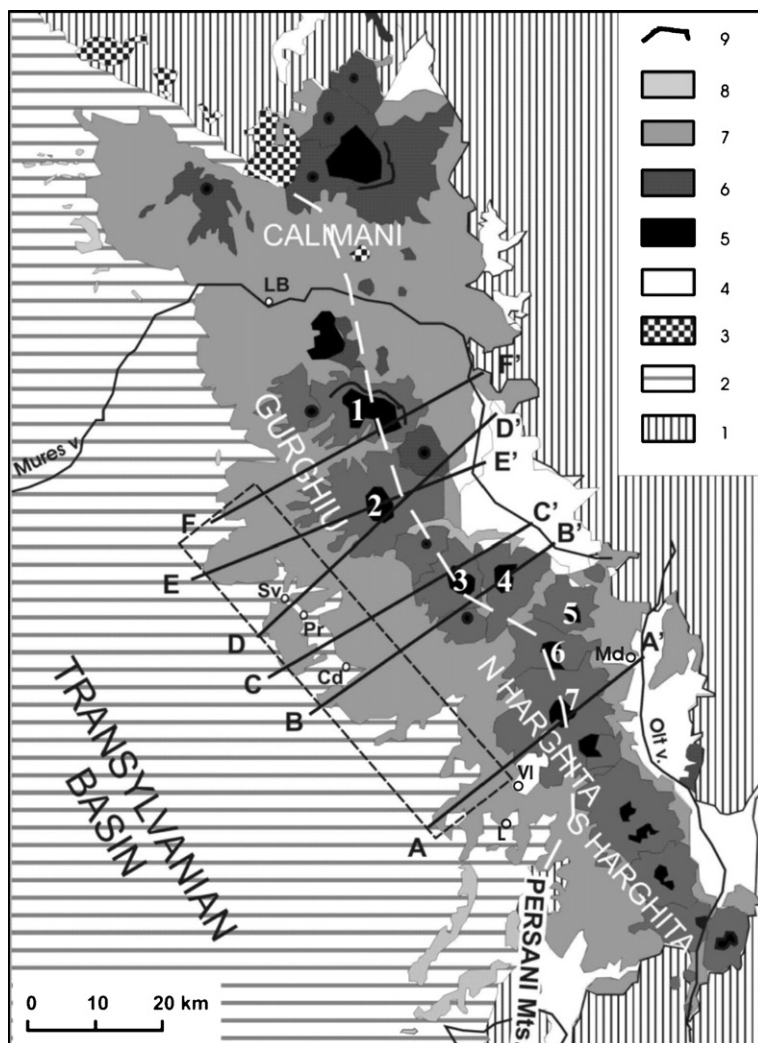
Fig. 2. Volcanic facies map of the CGH volcanic chain with location of morphological profiles (AA′–FF′). 1 — East Carpathian thrust-and-fold belt, 2 — Transylvanian Basin, 3 — Pre-volcanic intrusions, 4 — Upper Pliocene/Quaternary post-volcanic basins, 5 — Central volcanic facies, 6 — Proximal volcanic (cone) facies, 7 — Medial volcanic (volcaniclastic) facies, 8 — Distal m (alluvial) volcanic facies, 9 — Topographic caldera rim. Dashed white line shows the presumed boundary between Carpathian and Transylvanian pre-volcanic basement types, respectively. Solid black lines labeled A to F show locations of topographic profiles in Fig. 3. Dashed frame shows the rough outline of the area with inward-dipping peripheral structural surface (see explanation in text). Localities: LB — Lunca Bradului, Sv — Sovata, Pr — Praid, Cd — Corund, Vl — Vlăhița, L — Lueta, Md — Mădăraș. Volcanoes mentioned in text (indicated by white numbers): 1 — Fâncel-Lăpușna, 2 — Seaca-Tătarca, 3 — Șumuleu, 4 — Ciumani-Fierăstraie, 5 — Ostoroș, 6 — Ivo-Cocoizaș, 7 — Vârghiș (after Szakács and Seghedi, 1995, modified).

gradually increasing upwards up to 35–40° (Francis, 1993; Davidson and De Silva, 2000). Volcanoes in the Gurghiu and North Harghita Mountains were active between 9–4 Ma (Pécskay et al., 1995). They show a rather attenuated morphology, and their outer slopes are typically low-angle (10–13°). Some of them vaguely suggest a former upward-concave profile (Șumuleu, Ciumani-Fierăstraie and Ostoroș) while others (Seaca-Tătarca, Ivo-Cocoizaș) have rather upward-convex shield-like profile (Szakács and Seghedi, 1995) as it is

obvious in profile D–D′ of Fig. 3. The latter category volcanoes are higher-volume than those belonging to the former one. Erosion may have played a role in attenuation of their original topographic profile, but its influence was quite modest: average erosional topographic lowering rate is ca. 30 m/Ma at the Carpathian Neogene volcanoes (Karátson, 1996). Also, the outer-slope angles could not be modified significantly by such low erosion rates since the end of volcanic activity. We suggest that the present-day low-angle outer slopes and
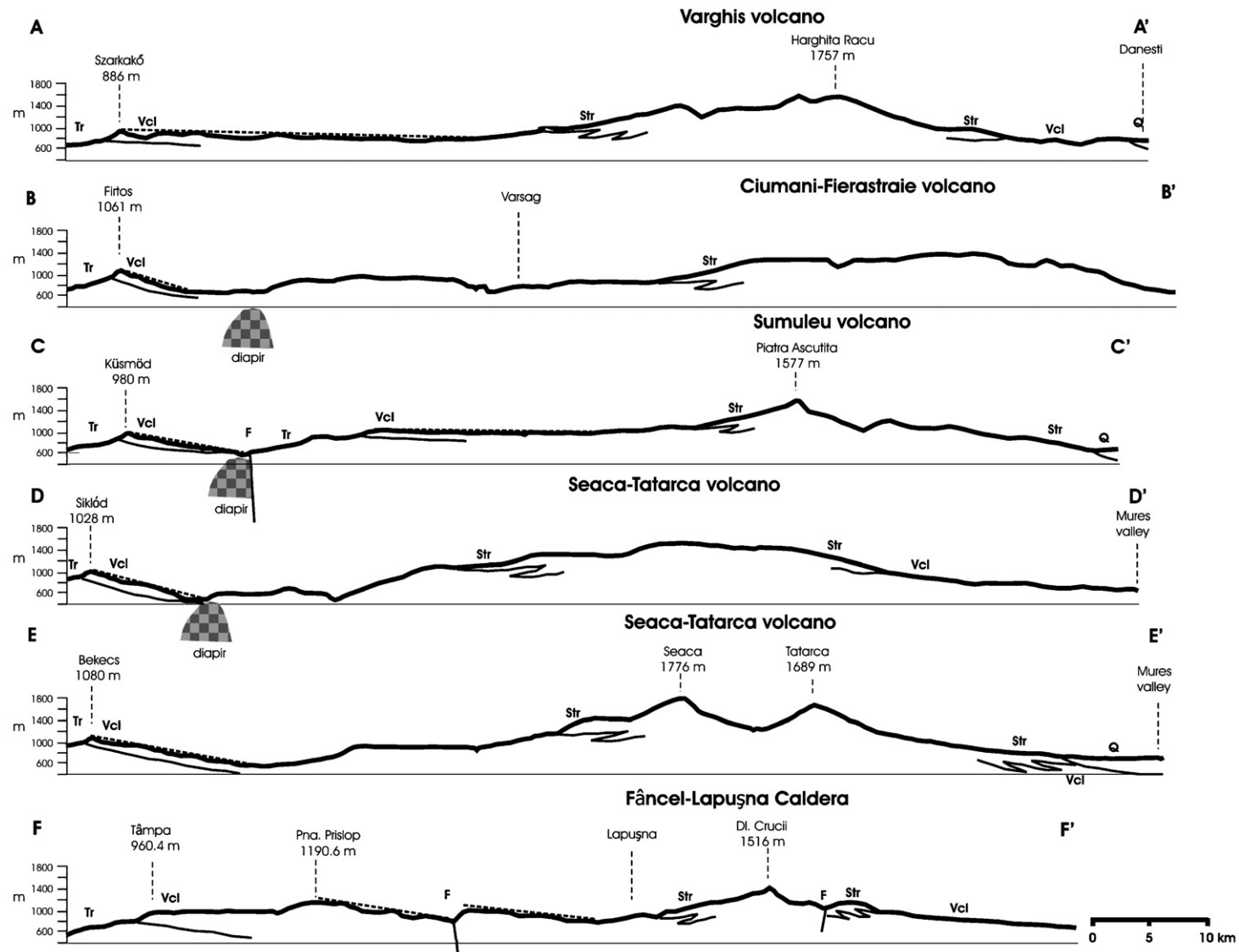
Fig. 3. Topographic profiles across the CGH (Gurghiu and North Harghita Mountains). Location of profiles in Fig. 2. Surface limits between main formations are shown. Tr — basin fill sediments of the Transvlvanian Basin, Vcl — peripheral volcaniclastics (medial volcanic facies), Str — proximal and central volcanic facies, Q — Quaternary sediments, F — faults. Fine dashed line shows smoothed topographic surface of peripheral volcaniclastic plateau. Note dip of peripheral topographic surfaces toward center of volcanoes and dip of basement–volcanics boundary in the same direction.
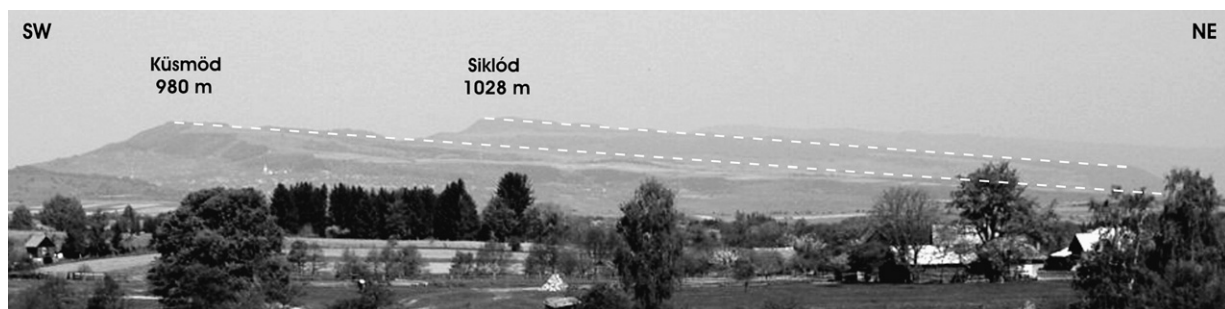
Fig. 4. Structural surfaces of peripheral volcaniclastic deposits in the western part of the CGH. Picture taken towards the northwest from south of Corund. Summit of Seaca-Tătarca volcano is at right, just out of sight of picture. Note dip of structural surfaces toward center of volcano shown by dashed lines. Slope break is visible on the opposite side of prominent hills marking the basement–volcanics boundary.

the shield-like morphology of the largest Gurghiu and North Harghita volcanoes, as well as the inward-dipping paleotopographic surfaces at the western peripheries of CGH chain are morphological features consistent with each other and may have resulted from post-eruptive subsidence of the edifices.

4.1.2. Geological evidence

The boundary between post-salt sedimentary basement and volcanics, seen in outcrops or inferred through mapping at the western peripheries of the Călimani, Gurghiu and North Harghita Mountains is invariably inward dipping, paralleling the paleotopographic surface of the peripheral volcaniclastic apron. This is especially well seen in the Sovata-Parid-Corund area at the western periphery of the Gurghiu Mountains (Figs. 3 and 4).

The central parts of most large composite volcanoes of CGH have been drilled but, none of them reached the pre-volcanic basement. Local networks of medium-depth drillings (150–600 m) were made to explore for iron ore deposits close to the western (Vlăhiţa-Lueta) and eastern (Mădăraş) peripheries of the North Harghita Mountains Since the siderite deposits are hosted at the basement–volcanics boundary, the paleomorphology of basement has additionally been investigated by gravity survey in the case of the Vlahiţa-Lueta area. Both drill-hole data and gravity survey show a general inward-dipping paleosurface of the pre-volcanic basement in the Vlăhiţa-Lueta area (Peltz et al., 1983), as well as at Mădăraş (Tănăsescu, 1967).

The obvious deeper position of the pre-volcanic basement surface in the axial part of the CGH chain with respect its peripheral areas has implicitly been interpreted in terms of volcano-tectonic depressions (Mureşan et al., 1986). The inferred "blind" normal faults explaining this subsidence were drawn on a purely theoretical basis without any positive evidence of their locations. Volcano-tectonic subsidence of basement beneath the axial part of the CGH cannot be ruled out. However, sagging of the volcanic edifices due to volcano spreading might be an equally viable explanation of central subsidence beneath the large composite volcanoes.

4.2. Volcano-induced tectonic features in the eastern Transylvanian Basin

A number of peculiar geological, structural and tectonic features along the eastern margin of the Transylvanian Basin are difficult to reconcile with the overall basin tectonics and evolution. They include contrasting tectonic style of the pre-salt and post-salt deposits, marginal thickening of the youngest basin-fill deposits, eastward dip of lithological and stratigraphic boundaries and of decollement surfaces in the post-salt sedimentary pile, presence of swarms of west-verging reverse faults, and enhanced salt diapirism. Such features can readily be identified on seismic profiles and by geological mapping.

4.2.1. Geological evidence

The youngest basin-fill sediments, Pannonian in age, occur along the eastern margin of the northern Tranylvanian basin. Their sedimentological features (Krézsek, 2005) indicate a relative high-angle Pannonian basin margin profile, characterized by coarse-grained fandeltas prograding on a narrow shelf, feeding sandy submarine fans (Fig. 5). Well-log data in the front of CGH chain indicate an overall eastward thickening of the Pannonian sedimentary pile toward the volcanic edifices. This is due to clockwise rotation — in vertical plan — of the Upper-Badenian to Pannonian succession along the salt layer acting as decollement surface. The

consequences of the large-scale rotation are: (1) the relatively flat Sarmatian/Pannonian depositional boundary was tilted eastwards (downward), (2) all Pannonian sediments were eroded in the proximity of the eastern Transylvanian salt-diapir belt; and (3) progressively younger sediments (Upper-Badenian to Pannonian) crop out toward the CGH.

4.2.2. Geophysical evidence: 2D seismics interpretation

A number of 2D seismic profiles have been acquired by oil companies during the 1990s in the Transylvanian Basin. A few of them reach the boundary area between the Transylvanian Basin and the CGH Chain (Fig. 6).

The seismic facies of the Upper-Badenian to Pannonian sediments were calibrated to the well-log facies and core data. The Upper-Badenian sediments are coarse-grained sands and conglomerates interbedded with small amounts of pelites. For the basal part of the succession, a proximal littoral environment is inferred, which is passing upwards into proximal fan deposits, built up by a complex fringing of channels and proximal

lobes. The Sarmatian is built up by various submarine fans. The change of seismic facies is mainly related to depositional changes characteristic for such an environment: superimposed submarine lobes and channels. The same depositional settings are inferred for the Pannonian deposits. The Upper Sarmatian and Pannonian seem to follow each other without erosional unconformities. It is important to note that the entire Upper-Sarmatian–Pannonian-volcanics succession is formed by relatively conformable, parallel reflectors. Therefore, we interpret stable sedimentary conditions, with almost negligible salt-tectonic activity. On the flattened seismic line (Fig. 7c), the Upper Badenian to Pannonian strata are onlapping the thinned salt-layer. However, the actual architecture indicates that the original onlaps were rotated downward. This rotation is due to the westward withdrawal of the salt layer beneath the onlapping deposits. If we suppose progressive salt withdrawal, than we must have angular unconformities between the subsequent Sarmatian seismic packages. However, this is not the case, rather it seems that the whole structure glided as a whole along the salt layer acting as a
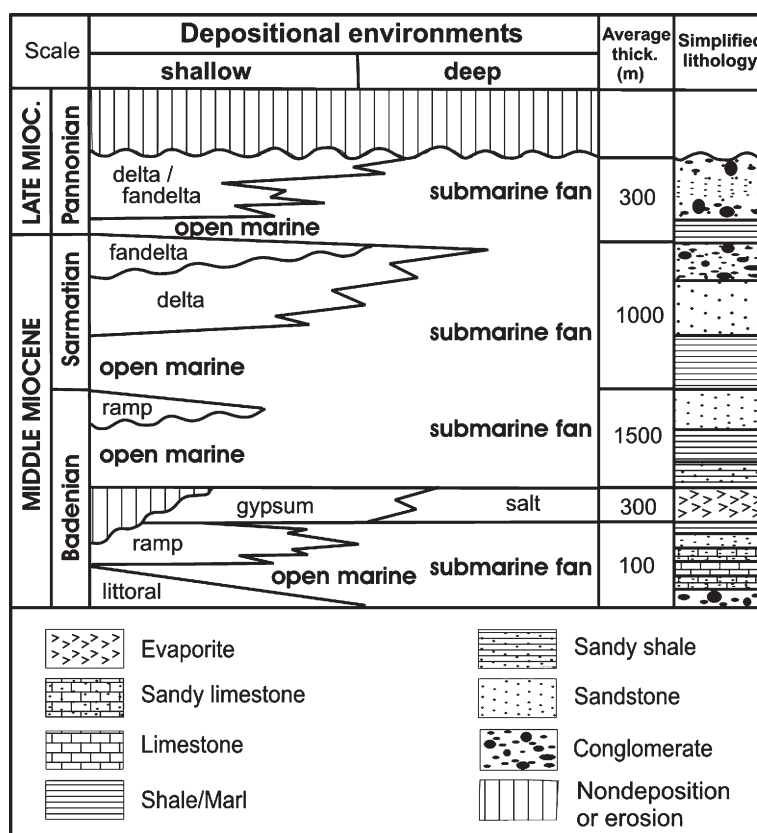

Fig. 5. Badenian to Pannonian sedimentary environments and simplified lithology of the Transylvanian Basin.

detachment surface after forming the volcanic edifices, in post-Pannonian times.

4.2.3. Enhanced salt tectonics

The Middle Badenian evaporites in the Transylvanian Basin are mostly represented by salt, forming a nearly continuous layer with variable thickness due to depositional conditions, doming and diapirism. Usually their thickness ranges between 100–300 m, but in the diapir structures it may attain up to 1500 m (Ciupagea et al., 1970; Visarion et al., 1976).

Although salt-dome structures are present throughout the whole Transylvanian Basin, outcropping and shallow diapirs are mostly concentrated along belts located near, and paralleling, the margins of the basin (Figs. 6 and 7a). One striking feature of salt diapirism in the basin is that the diapir structures are much better developed, involving much larger volumes of salt, in the eastern part as compared to the western part. This feature has not been explained so far.

Near-surface and outcropping salt diapirs in the eastern part of the Transylvanian Basin form a diapir belt roughly parallel to the basin margin and to the CGH volcanic chain. The northern half of this belt is located west of the western margins of the CGH volcanics (Călimani and northern Gurghiu Mountains), while its southern half is inside the medial to distal-facies volcaniclastic apron of the volcanic chain (southern Gurghiu and North Harghita Mountains) (Fig. 6).

Salt diapir structures can easily be identified and described on seismic profiles. The diapirs deform all the surrounding sediments, including the youngest Pannonian ones too. In profile A–B (Fig. 7a), crossing the whole basin, it is obvious that diapir structures in the eastern part are much larger (radius up to 3 km, and hight up to 3 km) than those from the western margin (height up to 1 km and width around 1 km). These observations lead to the conclusion that diapirism in the eastern part of the basin is strongly enhanced as compared to diapirism in the western part: both salt volumes involved and duration of the process show larger values. This requires the admission of additional factors involved in salt diapirism in the eastern part of the basin. In our opinion these could be: 1) the proximity of the Neogene CGH volcanic chain, 2) the position of the Middle Badenian depocenters eastward to the actual basin center, involving higher amounts of salt deposited here, and 3) the post-Pannonian uplift of the eastern margin of the Transylvanian Basin (Krézsek, 2004).
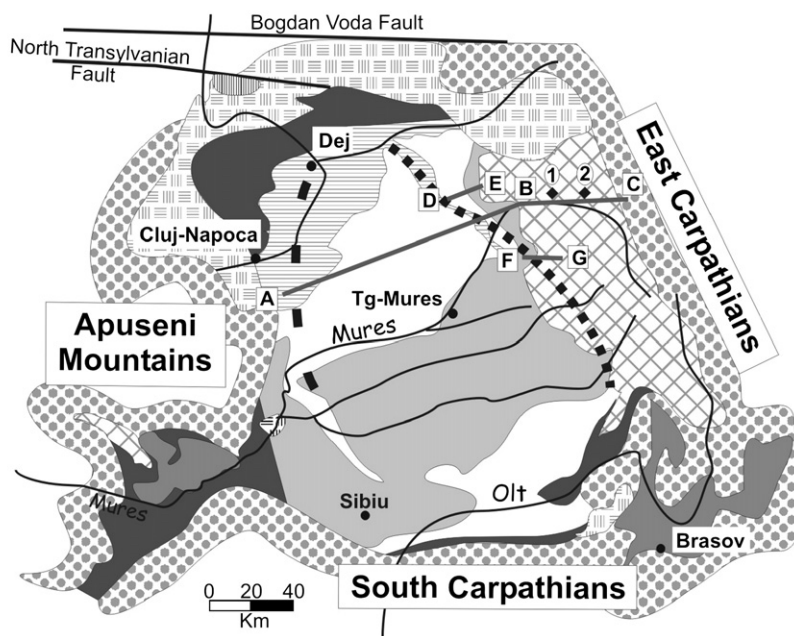
Fig. 6. Geologic sketch map of the Transylvanian Basin (compiled after Ciupagea et al., 1970) with location of 2D travel-time seismic profiles (A–B, D–E, F–G).
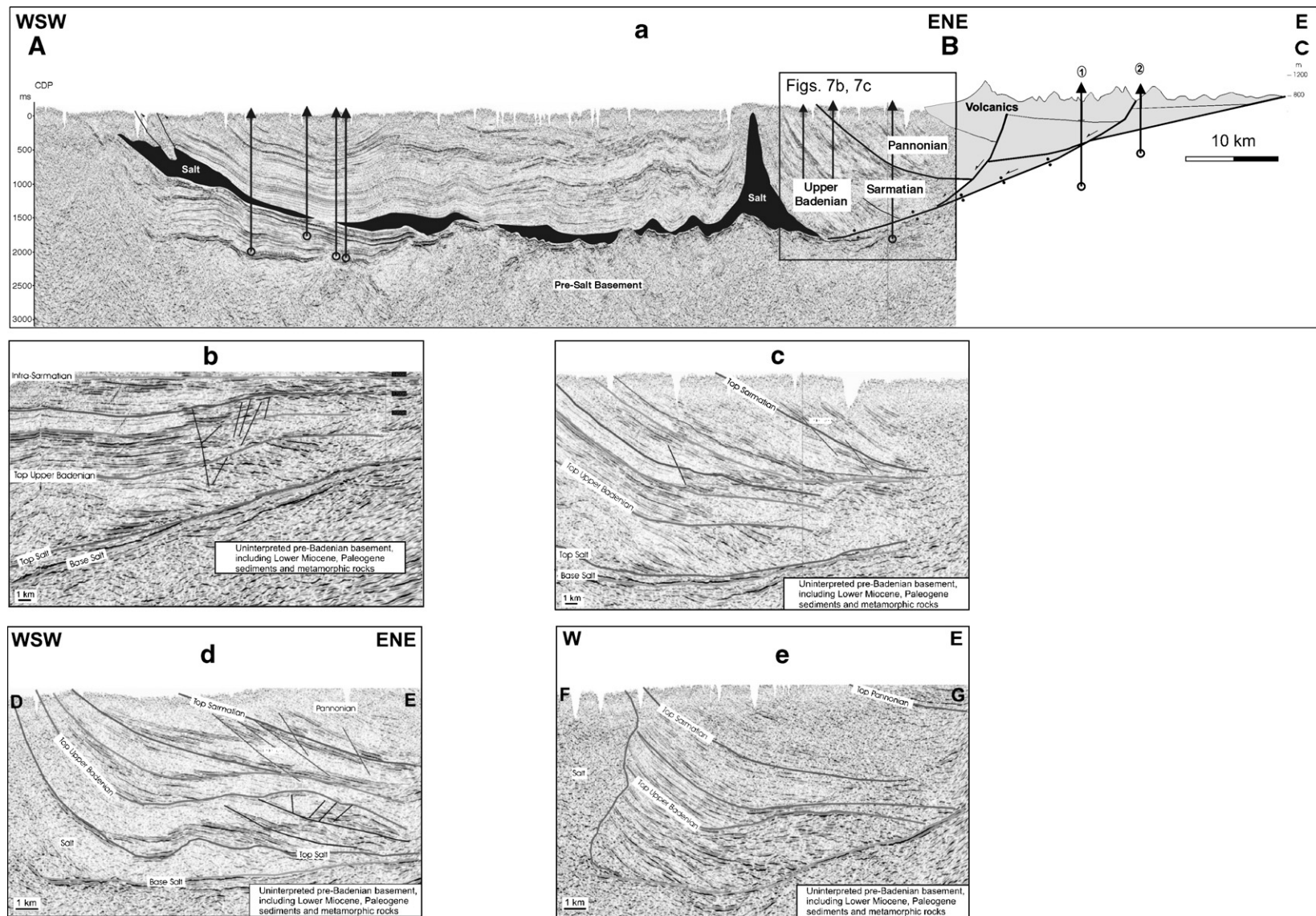
Fig. 7. 2D travel-time seismic profiles in the Transylvanian Basin. Locations are shown in Fig. 6. a. Interpreted regional seismic line (A–B) and its interpreted eastward extension (B–C). The tentative interpretation of the extended segment is based on outcrop and drill hole data. b. Interpreted detail of the regional seismic line A–B (a). c. Interpreted detail of the seismic line A–B (b), flattened on an intra-Sarmatian horizon. d. Interpreted detail of the regional seismic line D–E. e. Interpreted detail of the regional seismic line F–G.

Although the diapirs seen in profiles A–B and C–D (Fig. 7b, d) are vertical, Visarion et al. (1976) state that most of the diapirs are southwest-verging and crosscut by vertical faults. In their salt isopach map, Visarion et al. (1976) depict a salt-wall line as a west-verging thrust fault along the diapir occurrences in the eastern Transylvanian Basin. Ciupagea et al. (1970) also mention west-vergence of the Miocene sediments including salt diapirs along the eastern margin of the Transylvanian Basin. Incipient small-scale diapiric features located east of the main diapir in profile A–B clearly show eastward-dipping axes (Fig. 7d). West verging of diapir structures is fully consistent with observations related to the westward gliding of the post-salt succession along the salt layer acting as a decollement surface. The gliding induced shortening near the Pre-Pliocene diapir belt, shifting the diapir axis westward.

4.2.4. Reverse faults

West-verging post-Middle Miocene reverse faults were previously pointed out in the eastern part of seismic line 1 (Ciulavu, 1999). They are located between the salt diapir belt of the northeastern Transylvanian Basin and the Călimani segment of the CGH volcanic chain. These faults deform only the Miocene sediments above the Middle Badenian salt layer and they display reverse vergence with respect thrust-faults occurring in the basin basement, shown in seismic line C–D (Fig. 7d).

Sanders (1999) and Huismans (1999) suggest that the reverse faulting in the eastern Transylvanian Basin is correlated with E–W compression in the East Carpathians because their orientation is similar to that of the back-thrusts in the internal parts of the East Carpathians. However, this interpretation ignores the fact that these reverse faults only affect post-salt deposits and no important west-verging thrust faults are recorded on the contact zone between the Transylvanian Basin and Middle Dacides (Balintoni and Petrescu, 2002).

The presence of west-verging reverse faults in the area reflect smaller-scale ruptural tectonics (thrust faults) related to the gliding of the post-salt brittle-lithology succession. Outer-fan to basin-plain Sarmatian to Pannonian pelites favorised the formation of local detachment surfaces along which the gliding-induced shortening was accommodated (Fig. 7b).

5. Tectonic interpretation

A narrow band of contractional deformation can be outlined along the eastern part of the Transylvanian Basin which is parallel to the volcanic chain, as well as to the alignment of salt-diapir structures. Moreover, it is parallel to the strip of inward-dipping structural surfaces of the volcaniclastic plateau.

The present structural architecture of the post-salt succession in the eastern Transylvanian Basin is related to salt-tectonics and salt-withdrawal processes, enhanced by volcano-spreading and regional tectonics. We found the existence of at least two important phases of salt-tectonic activity: one around the Badenian–Sarmatian boundary, and the other post-Pannonian. The first phase is related to one of the pulses of the compressive tectonics of the Carpathian Chain. Following the Late Badenian salt-tectonic phase, probably due to the high rate of sedimentary input, beneath the Upper-Badenian–Sarmatian to Pannonian sediments a relatively thick salt-layer was trapped. It was subjected to volcano-induced tectonics during the second salt-tectonic maximum, which postdates in time the building up of the large volcanic edifices in the central CGH. The effect of volcanism in enhancement of diapirism is twofold: (1) increased geothermal gradient resulted in progressively decreasing salt viscosity near the CGH range, and (2) compressive stress on the pre-volcanic basement due to gravitational loading by volcanic edifices. The resulting stress-field has two components: a downward-oriented component resulting from the weight of volcanic edifices and a westward component induced by the preferentially westward spreading of volcanic structures due to both westward-sloping substratum and buttressing in the east.

There are major structures in the Transylvanian Basin related to the Upper-Pannonian to Pliocene paroxysmal tectonic phase. It is striking that the salt-withdrawal process and related tectonics seem to be of the same age. Therefore, we suspect that the triggering mechanism of the post-Pannonian salt tectonics was represented by this regional tectonic phase. The salt tectonics was enhanced by the preferential westward spreading of the volcanic edifices, as well as by volcanism-related increased heat flux. The post-Pannonian salt-tectonics represented a major salt-withdrawal process generating large-scale detachment along the remaining thinned salt layer. The post-salt succession, including the volcanic edifices, was displaced downward, following the westward migration of the salt. This downward displacement created a westward-oriented compressive stress-field, pushing the post-salt sedimentary succession onto the progressively evolving eastern diapir belt. At the end of the process, reverse detachment listric faults have been created above the plastic salt-layer, while the post-salt succession siliciclastic pile was pushed upwards. The west-verging

thrust faults thus represent small-scale compression structures when the block-rotation and volcanic spreading-induced stress have been locally compensated. They do not account for the large-scale architecture of the present structures.

6. Discussion of volcano–basement interaction

Basement is different beneath the eastern and western parts of the CGH volcanic chain. Proterozoic to Lower Paleozoic metamorphic rocks overlain by Mesozoic mostly carbonatic sedimentary rocks (belonging to the Inner Dacides Units of the East Carpathians, Sǎndulescu, 1988) form the pre-volcanic basement in the eastern part. Due to this composition, its mechanical properties allow for brittle deformation under stress. In contrast, the pre-volcanic basement of the western part of the CGH consists of a thick sedimentary pile including ductile rocks such as clay and salt prone to plastic deformation. The boundary between these two contrasting basement types is located beneath the axial part of the CGH, excepting its southeasternmost segment where the row of South Harghita volcanic edifices crosscut the Inner Dacides (Szakács et al., 1993). The geometric details of this major structural boundary are as yet unknown.

Sagging of the central composite volcanoes into such a compound basement with contrasting mechanical behavior should be non-uniform. We may tentatively assume that basement of the eastern part deforms by either flexuring or faulting under the weight of the volcanoes. Such a basement-deformation style should be restricted to the axial part of the CGH and deformational structures do not propagate peripherally, being hidden beneath the thick volcanic pile.

Volcano–basement interaction in the western part of the CGH is much different. The presence of plastically deformable rocks in the basement may induce stronger and more complex deformation of underlying basement under the weight of the volcanoes, while the edifices themselves may also deform. Space problems caused by sagging of the volcanic edifices are resolved by the mechanisms of volcano spreading (Delaney, 1992; Borgia, 1994) which may affect the basement only or both basement and edifice, as suggested by scaled experiments (Merle and Borgia, 1996). In the case of the CGH chain and its western-half basement the signatures of both processes can be recognized in the geologic and topographic record as discussed in the previous sections. Sagging of the central part of the volcanic edifices led to inversion of dip direction of the volcanic rocks–

basement boundary at the western periphery of the CGH chain, as well as of the paleotopographic surface of the peripheral volcaniclastic plateau (change of topographic gradient from outward-dipping to inward-dipping).

Sagging also resulted in squeezing of plastic rocks (especially salt) from beneath the volcanic cones. At isolated volcanoes such a process may generate deformation structures in basement rocks all around the volcano in a concentric arrangement (Merle and Borgia, 1996). In the particular case of the CGH edifices, compressive deformation of basement rocks is strongly focused to the west, to a sector of ca. 1/4 of the volcanoes circumference. Focused deformation of basement at preferential sectors is due to the following factors: (1) buttressing by plastically undeformable basement in the eastern half of CGH, (2) westward dipping basement topography; and (3) close spacing of adjacent composite volcanoes, which does not allow for spreading along the CGH chain axis. Thus, volcano-induced compressive stress in plastic basement rocks is constrained to act unidirectionally, normal to the chain axis. As a consequence, the resulting deformation is expected to be stronger and farther-reaching as compared to cases where stress is more uniformly distributed around volcanoes. Inward-dipping paleotopographic surfaces and volcanic rock–basement boundaries are located at ca. 22–28 km from the geometrical axis of the CGH. Reverse fault swarms seen in seismic sections are found at up to 30–35 km distances. These figures are significantly larger as compared with most of those reported elsewhere. The volcano-induced morphological Alejuela Ridge, for instance, developed above a thrust-fault, is at ca. 14 km from the Central Costa Rica Volcanic Range (Borgia et al., 1990). At Nicaraguan volcanoes Concepcion and Maderas, basement-deformation features including thrusting and mudstone diapirism are within 10 km from their summits (Van Wyk De Vries and Borgia, 1996). However, fault-propagation folds around the base of the Etna volcano, interpreted as resulting from volcano–basement interaction, are found up to ca. 30 km from summit (Merle and Borgia, 1996), but Etna's edifice is also buttressed (on two sides: north and west). The influence of inclined basement topography on volcano spreading and edifice failure, as recently suggested from analogue modeling (Wooller et al., 2004), acts in the downslope sense of sectorial deformation in the direction of basement tilt.

The styles of volcano-induced deformation have been addressed through numerical modeling by Van
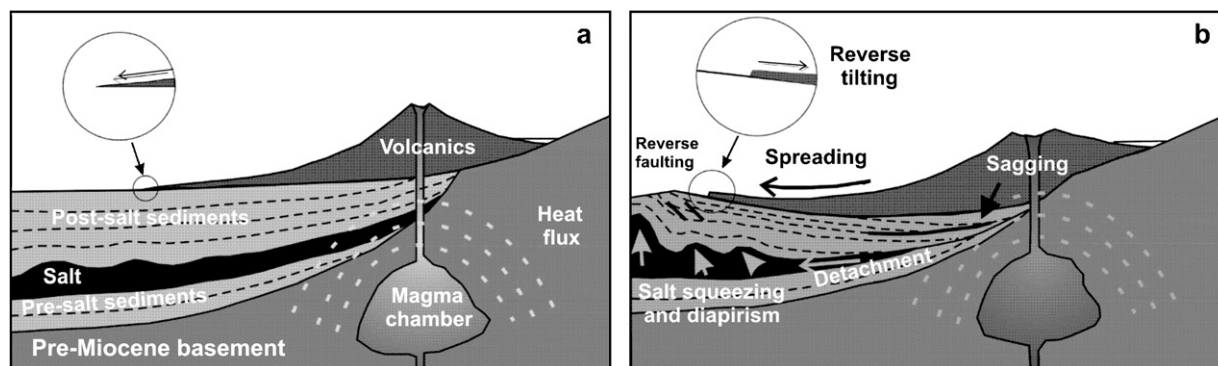
Fig. 8. Cartoons showing a model of geological evolution with snapshots before (a) and after (b) volcano–basement interaction in the eastern Transylvanian Basin. See text for discussion.

Wyk De Vries and Matela (1998). According to numerical parameters — the ratios volcano radius/ductile layer thickness (Π_a) and viscosity of ductile substratum/failure strength of volcano (Π_b) — they identified basement-deformation regimes including flexure, spreading and extrusion. In our study case, salt is the ductile layer. Although clay is also present in the Sarmatian–Pannonian sedimentary pile above the salt, it is interbedded with sandstone and the resulting sequence (of which ca. 50% is clay) is far less ductile than salt. Since salt thickness is 100–300 m, in the Transylvanian Basin, excepting the diapirs (Visarion et al., 1976), the ductile layer is thin beneath the CGH volcanoes. The salt is 71 m thick at Lunca Bradului (Mureş valley, Fig. 2), well inside the CGH volcanic chain, as drill-hole data show (Ciupagea et al., 1970). The deformational features found at the western periphery of CGH, both in basement and the volcaniclastic plateau seem consistent with the thin ductile layer model of Van Wyk De Vries and Matela (1998) in which the volcano and its shallow rigid substratum are coupled and deformed together by lateral spreading facilitated by the ductile salt layer beneath. Sagging of the central part of volcanoes can also be viewed as a result of extensional stress related to the lateral spreading of basement, in addition to subsidence due to squeezing out of the ductile rocks from beneath. Although this model is consistent with most observations, it is unlikely that it applies in a pure form because it assumes high viscosity of the plastic layer while salt acts as a low-viscosity medium. Lateral extrusion of salt may have played an important role too. Enhanced diapirism in the eastern Transylvanian Basin as compared with that in its western part involves higher volumes of salt available and mobilized. Westward extrusion of salt from beneath the volcanoes under their

load may account for this extra volume. Salt viscosity, plasticity and flowage could be significantly influenced by thermal effects due to higher heat-flow in the proximity of volcano-related heat sources (magma chambers). Temperatures of 50 °C have been measured in artesian brines in a drilling near the salt dome at Praid (Ciupagea et al., 1970), suggesting heat-flow anomalies in the area. The possible contribution of volcano-generated heat to substratum plasticity and, thus, to volcano spreading, has not been addressed yet in the relevant literature.

In summary, a model involving both basement spreading and extrusion due to the presence of a thin ductile layer in the substratum seems to us a reasonable scenario of volcano–basement interaction in the eastern Transylvanian Basin (Fig. 8 a, b). It involves only the western part of the CGH volcanoes, where ductile layer is present in their substratum. However, volcano-induced tectonics should not be considered alone as accounting for the complexity of structural features recorded in the post-Paleogene basement of CGH. Rather, volcano–basement interaction developed on a background of intense regional tectonic activity, and the resulting features reflect the interplay between regional and local stress fields in a complex way.

Acknowledgements

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