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GEOMORPHOMETRIC ANALYSIS AND THE EVOLUTION OF DRAINAGE NETWORK IN TRASCĂU MOUNTAINS (ROMANIA)

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Abstract: Drainage network evolution of the Trascău Mts, including the formation of the famous gorges (e.g. Cheile Turzii) is constrained by the transitional position between the high central Apuseni Mts and the Mureş Valley, further on by the ~N-S oriented geologic settings. The aim of this study is to use digital terrain analysis tools in order to better understand this evolution. Elevation, slope and aspect distributions, topographic swath profiles, stream profiles and doline morphometry were analyzed based on the SRTM dataset, topographic maps, and partially on field GPS measurements. Elevation and slope histograms according to rock groups quantitatively supported differences in rock resistance. It is demonstrated that Cenozoic rocks can be well distinguished from Mesozoic and older rocks based solely on morphometric parameters (slope and standard deviation). Swath analysis highlighted a characteristic W-E change in the slope of the envelope surface that is attributed to tectonic movements. Swaths profiles also helped the recognition of water gaps and wind gaps, which are very important remnant landforms of the post-Cretaceous drainage network. Stream profiles of the study area can be modelled mostly by exponential and linear functions, and a large number of identified knickpoints are in relation with rock boundaries. The denudation blocking effect of the main limestone ridge is clearly seen on stream profiles. It is demonstrated how the original Post-Cretaceous radial drainage pattern evolved to a trellis pattern. It is argued that superposition (with antecedence) played the most important role in the formation of water and wind gaps. All analysis highlighted the differences between the areas north and south of Arieş river. These areas had similar landform evolution, but are at different stages. The Post-Sarmatian evolution of the northern part copies the Post-Cretaceous evolution of the southern part. Differences in doline density between the northern and southern parts are attributed to different duration of subaerial karstification.

Keywords: DEM, geomorphometry, swath profile, drainage network, stream profile, karst, Trascău Mountains, Apuseni, Cheile Turzii

1. INTRODUCTION

The Trascău Mts are located in the eastern part of Apuseni Mts in a transitional position between the ~1800m (a.s.l.) high „core" of the Apuseni Mts (Gilăului Mts) and the Transylvanian basin, namely the Mureş Valley at ~250m (a.s.l.) as the local erosion base (Fig. 1A). This transitional position and the variegated geological settings constrain the geomorphology of the Trascău Mts. The most remarkable geomorphological features of the area are the deep gorges (e.g. Cheile Turzii, Turului, Râmeţului, Intregalde) crossing the NNE-SSW oriented narrow limestone ridges. Earlier geomorphic studies mainly focussed on the formation of these gorges (e.g. Martonne, 1922; Cholnoky, 1926; Nyárádi Erasmus, 1937; Tulogdi, 1943; Cocean, 1988), the question of denudation surfaces (Martonne, 1922; Ficheux, 1937; Cocean, 1985) or gave a complex geomorphologic description (Popescu-Argeşel, 1977). However, recent geomorphologic literature is not so abundant (e.g. Korodi, 2006) and no detailed GIS-analysis has been performed for the Trascău Mts until now. Therefore, the aim of our study is to carry out a digital elevation model (DEM) based terrain analysis for the area and to present how geomorphometrical tools may help to understand and quantify landform evolution in a transitional mountain range.

2. GEOLOGIC-TECTONIC SETTINGS AND PREVIOUS STUDIES ON THE DRAINAGE NETWORK EVOLUTION

The geologic settings are shortly summarized after Giuşcă & Bleahu (1967; Fig. 1B). Metamorphic rocks (micaschists, paragneiss, metavolcanic rocks and crystalline limestone) are present in the western and north-central part of Trascău Mts. The landscape is dominated by the Upper Jurassic Stramberg Limestone. Basaltic ophiolites of the same age are closely connected to the limestone ridges. In the central and southern parts, these Jurassic and metamorphic rocks are surrounded by Cretaceous sediments (conglomerate, sandstone, marl, argillite). Huge limestone olistholits are found in the Lower Cretaceous flysch deposits. The northern and eastern parts of the study area are covered by Tertiary sediments, these are mainly Miocene marine sediments, but Paleogene sediments also outcrop around the Iara valley. Quaternary rocks are present as river terraces especially along the Mureş valley, and as alluvial sediments in the small intramontane basins (Iara, Rimetea, Sălciua). Volcanites are present in a negligible extent only, but granitic rocks have larger extension in the neighbouring Gilăului Mts.

Contrary to the geomorphological literature, recent geological and geophysical investigations are widespread. Most of them focuses on the Mesozoic structural formation of the Apuseni Mountains (e.g. Bălc et al., 2012; Kounov & Schmid, 2013) and on the Cenozoic evolution of the Transylvanian Basin including the surrounding mountains (e.g. Ciupagea et al., 1970; Ciulavu et al., 2000; Sanders et al., 2002).

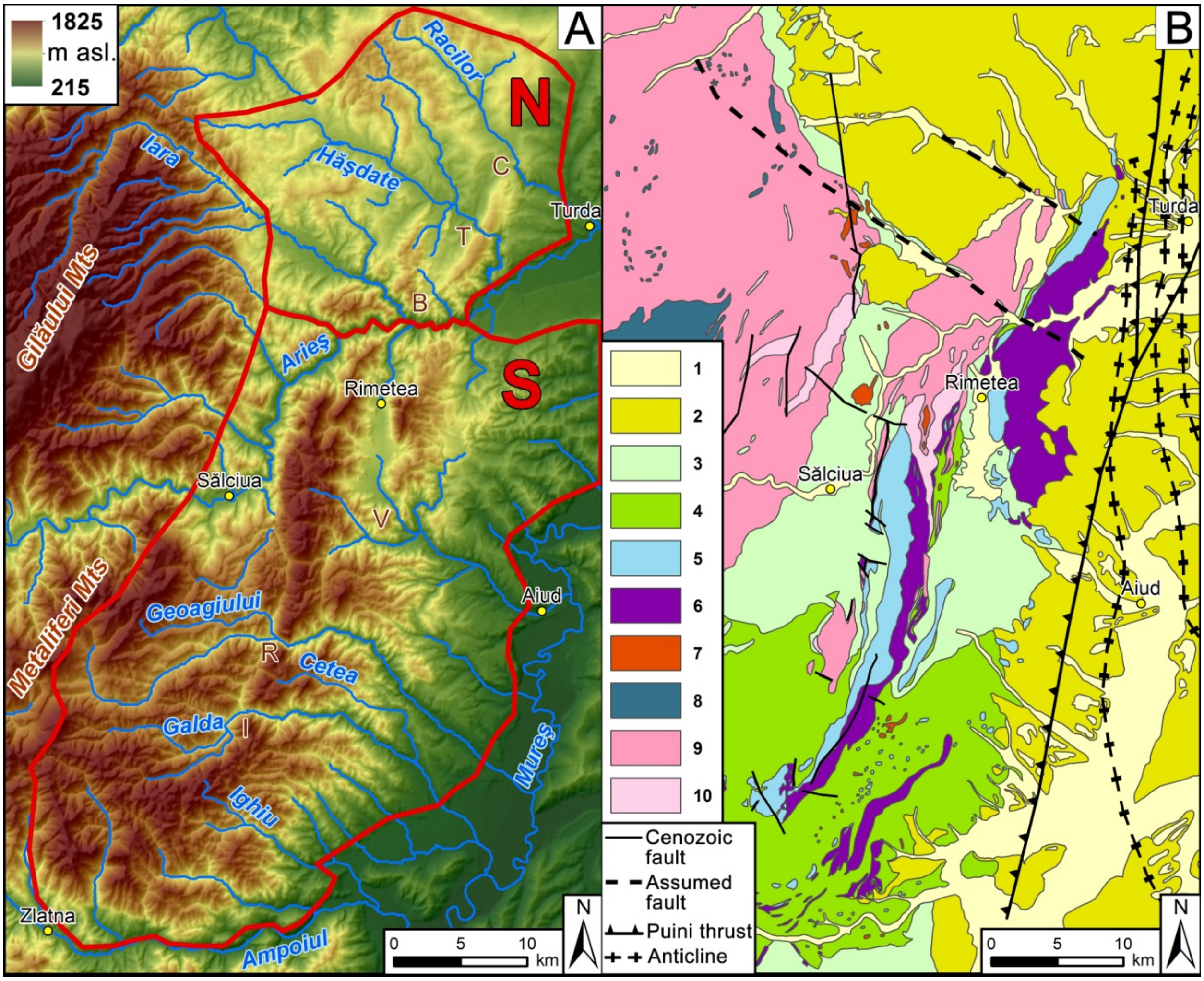

Figure 1. A) Elevation map of the study area (gorges C:Turului, T: Turzii, B: Buru, R:Râmeţului, I: Intregalde, V: Vălişoarei; analysis units N: northern, S: southern). B) Simplified geologic map with Cenozoic structural lines after Giuşcă & Bleahu (1967) and Kounov & Schmid (2013). 1: Quaternary Sediment, 2: Tertiary Sediment, 3: Upper Cretaceous Sediment, 4: Lower Cretaceous Sediment, 5: Jurassic Limestone, 6: Jurassic Ophiolite, 7: Tertiary Volcanite, 8: Palaeozoic and Precambrian Granite and Gneiss, 9: Metamorphic rocks, 10: Crystalline Limestone.

Several papers deal with the complex situation of the nappe stack (e.g. Bleahu et al., 1981, Balintoni & Iancu, 1986; Kounov & Schmid, 2013). Kounov & Schmid (2013) demonstrated that the nappe system was formed during 3 tectonic phases (Austrian, Turonian and Laramian) with thrusting towards East in the Early Cretaceous, then thrusting towards West in the Late Cretaceous. They also emphasized the importance of Puini thrust, which was first formed during the latest Cretaceous, but reactivated later in the Paleogene.

Development of the Transylvanian basin began in the Late Cretaceous (Ciulavu et al., 2000). It is widely accepted (e.g. Sanders et al., 2002; Kounov & Schmid, 2013) that the rapid subsidence of the Transylvanian Basin started in early mid-Miocene times. The Transylvanian basin has been surrounded by higher topography at all sides since the Late Badenian-Sarmatian (15-11 Ma) according to Ciupagea et al., (1970), Săndulescu (1988) and Sanders et al., (2002). The surrounding continental land surface was characterized by low and smooth landforms in the pre-Neogene (e.g. Ielenicz & Simoni, 2007). Since the Pliocene, the whole basin has been in uplift (Ciulavu et al., 2000). Sanders et al., (2002) pointed out that from Pliocene to recent there was an isostatic surface uplift of about 300-500 meters, and substantial vertical movements (>1000m) must have taken place after the Early Badenian. They calculated that the eroded material since the Badenian was at least 1000m for the eastern part of Apuseni Mts. Kercsmár et al., (2012) described two phases of uplift of the Apuseni Mountains during the Late Neogene and Quternary, the second of which was characterized by an erosion-driven, isostatic uplift.

As for the denudation surfaces, Ielenicz & Simoni (2007) summarized the actual knowledge about erosion surfaces in Romania. They described the following stages for their formations: the „Carpathian pediplain" developed from late Cretaceous to Eocene/Oligocene; the „medium Carpathian" and the „Carpathian border" surfaces, which were formed in the Miocene and Early Pliocene, and finally, erosion levels and glacises, which were formed during the Late Pliocene. In the Apuseni Mts and especially in the Trascău Mts, Martonne (1922), Ficheux (1937), Popescu-Argeşel (1977), Cocean (1985) and Móga (2002) described the erosion surfaces. There are three levels: the Ciumerna-Bedeleu (1000-1200m a.s.l.), the Râmeţ-Ponor (700-900m a.s.l.) and the Pliocene surface (4-500m a.s.l.).

A remarkable feature of the present drainage network, that the flow direction of rivers is usually rectangular to the orientation of the main geologic units. According to Popescu-Argeşel (1977), the rivers flowed towards north at both sides of the main limestone ridge in the Tortonian. Later on, due to the late Miocene-Pliocene subsidence of Mureş Valley, the eastern valleys flowing towards southeast, were formed by strong headward erosion and cut the limestone ridge at Cheile Râmeţului, at Cheile Intregalde and at Buru (Fig. 1) By the end of Pliocene, the river network was similar to the present one. However, we rather agree with Cocean (1988), who stated that headward erosion is not possible in case of limestone ridges. He explained the formation of Cheile Râmeţului by karstic capture, Cheile Intregalde by antecedence and Cheile Turzii, Turului and Vălişoarei by superposition.

3. METHODOLOGY

Our study area (Fig. 1) is somewhat larger than Trascău Mts, in order to better understand the drainage evolution. The boundaries of the study area were determined taking into consideration both geology (especially at the western boundary) and topography (drainage divide to the north, river valleys to the east and south). For the digital terrain analysis, we used the NASA SRTM (Shuttle Radar Topography Mission) database (for details see Rabus et al., 2003). Its 3" horizontal resolution is suitable for a mountain scale study. The DEM was reprojected to UTM coordinates. Slope values were calculated from the DEM. It is noted that slope values are slightly underestimated due to the medium horizontal resolution of the SRTM database (cf. Kienzle, 2004). Aspect values were calculated from the mean-filtered DEM (using 1500 m circle radius for filtering) in order to get the generalized aspect of the smoothed topography and to ignore the biasing effect of small-scale landforms. A standard deviation filter (with 1500m circle radius) was also applied for the DEM in order to characterize general surface dissection.

Elevation histograms were used to calculate the characteristic altitude range of certain terrains and the superficial extension of different rock groups. The vertical resolution of the histograms was 20m in all cases. We used swath profiles in order to evaluate surface trends. Swath profiles are generalized cross-sections, in which minimum, mean and maximum elevation values within a given swath versus distance are presented (Telbisz et al., 2013). The swath profiles are widely used in tectonic geomorphology, because in many cases the maximum curve represents the remnant surface, the mean curve illustrates the general trend of the surface smoothing the disturbing effects of small-scale landforms and the minimum shows the trend in the valley thalwegs.

Stream profiles were also analyzed, but their calculations were based on 1:25000 scale topographic maps, since SRTM DEM errors may cause significant errors in stream profiles (Eisam Eldeen & Telbisz, 2012). Knickpoints were automatically detected by calculating the slope changes in the stream profiles. Knickpoints may be caused by rock boundaries, stream confluences and tectonic movements. Whereas the location of the bedrock-related knickpoints is stable, the other type knickpoints gradually recede by headward erosion (e.g. Hack, 1973; Bishop et al., 2005; Goldrick & Bishop, 2007; Larue, 2008). Different mathematical functions (logarithmic, exponential, power and linear) were fitted to profile shapes, but usually for selected segments only and not for the whole profile, because of the compound nature of profiles.

In order to quantify differences in the intensity of karstification, doline morphometric data were calculated from 1:25000 scale topographic maps, but these data were completed by field GPS measurements as well.

4. RESULTS

4.1. Elevation, slope and aspect distributions

Since the areas north and south of Arieş river are so different from both geological and geomorphological point of view, we performed the analysis distinctly for the two parts (Fig. 1A). As for the elevation (Fig. 2), one may observe three maxima for the southern part: at 370-430 m a.s.l., at 510-550m a.s.l. and at around 890 m a.s.l. For the northern sector, there is a unique maximum at 570m a.s.l. The lowermost level is found at the eastern side of the mountains on the wide interfluvial ridges and it may be assigned to the Pliocene level though it is sligthly lower than the Pliocene surface mentioned in Section 2. The second level is attributed to the Rimetea basin, so it is formed by deposition and not by erosion. The somewhat higher characteristic elevation range in the northern part is linked to the large, smooth surface west of Cheile Turzii. The third level observed in the histogram is present in the southern part, east and west of the main limestone ridge. This level may be linked to the Râmeţ-Ponor surface, but it is somewhat higher than according to the aforementioned authors. The uppermost Ciumerna-Bedeleu level is not recognizable at all in the elevation histogram. However, if we recalculate the histogram for terrains where slope is less than 10°, then the lower maxima are preserved and even emphasized, whereas a new, small maximum is also observable at 1210m a.s.l.

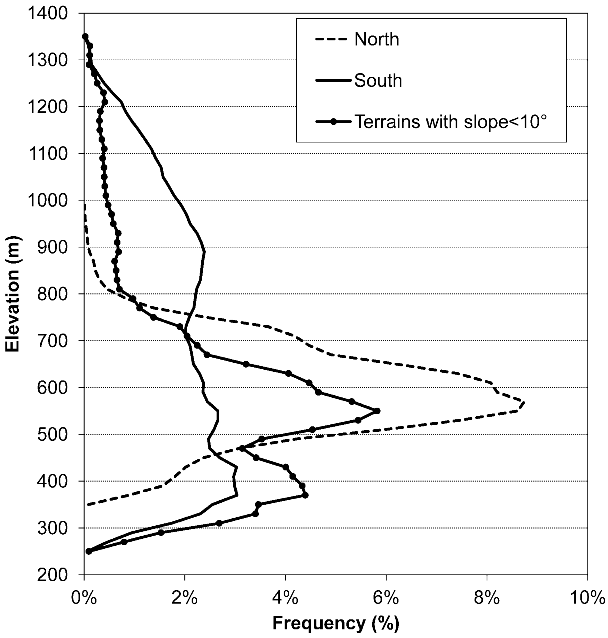

Figure 2. Elevation histograms.

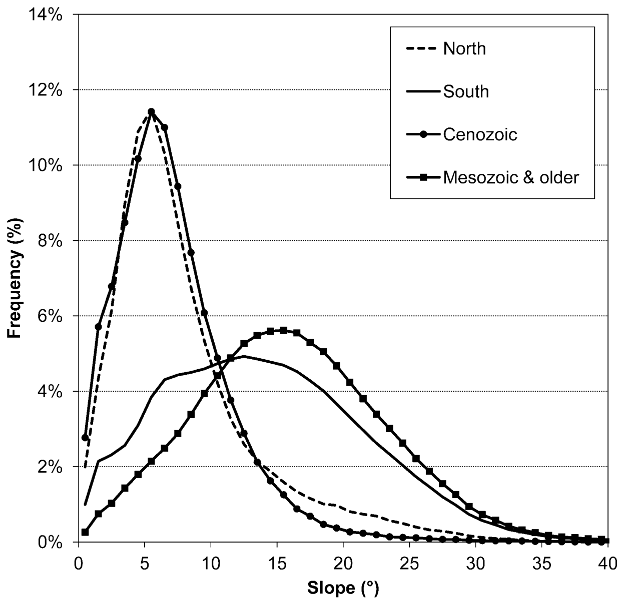

Figure 3. Slope histograms.

Taking into consideration the slope distribution (Fig. 3), the difference between the northern and southern parts is also significant. The most frequent slope category for the northern terrain is 5-6°, whereas it is 12-13° for the southern part. This difference in the characteristic slope range is predominantly the result of the dissimilar geological composition since the northern area is mostly covered by Tertiary and Quaternary rocks, whereas the southern part is characterized by Mesozoic and older rocks.

Given the fact that the area was a gently undulating terrain as a result of the Carpathian pediplain development, the Mesozoic and older rocks were found in a relatively narrow elevation range by the end of Paleogene. Therefore, the present-day differences in the elevation ranges of rocks may be caused either by vertical tectonic movements or by differential erosion. The same is true for Miocene marine sediments, which also had to be at similar elevations at the time of their formation. That's why we compared the elevation and slope histograms of the present surface extent of different rock groups (Fig. 4).

First, it is observed that Quaternary and Tertiary rocks are at much higher elevations in the northern area than in the southern, namely, the most frequent elevation of Quaternary rocks is at 530 m a.s.l. in the north against 330 m a.s.l. in the south, and the frequency maximum of Tertiary rocks is at 570 m a.s.l. in the north against 370 m a.s.l. in the south. For Mesozoic and older rocks, the elevations are higher in the southern terrain. In the southern part, the Jurassic Limestone has by far the highest maximum at 1110m a.s.l. The crystalline limestone is also found in this elevation range, but its maximum is at 1000m a.s.l. After the limestones, the order is Lower Cretaceous (900m a.s.l.), Jurassic Ophiolite (770m a.s.l.), metamorphic rocks (730m a.s.l.) and Upper Cretaceous (600 m a.s.l.). This order is partly due to rock resistance, but partly due to drainage evolution (see later). In the northern part, the highest maximum is also linked to the Jurassic Limestone (650m a.s.l.), but Metamorphic rocks are found higher than Jurassic Ophiolites.

As for the slope distribution (Fig. 3), there is a remarkable difference between Cenozoic and pre-Cenozoic rocks. It is calculated that 79.5% of the area covered by Cenozoic rocks has less than 10° slope angle. On the contrary, 79.9% of the area covered by Mesozoic and older rocks has slope angle higher than 10°. The most frequent slope category of the Cenozoic rocks is 5-6°, while it is 15-16° for the older rocks. Based on a slope map, the boundary between Cenozoic and pre-Cenozoic rocks coincides quite well with the 12° slope isoline. An even better morphometric determination of this boundary can be achieved by applying a standard deviation filter (with 1500m filter radius) for the DEM. This way, the agreement between the 60 m standard deviation isoline and the Cenozoic boundary is very good (Fig. 7A). Surface aspect may give useful information, if consequent valley direction is to be determined.

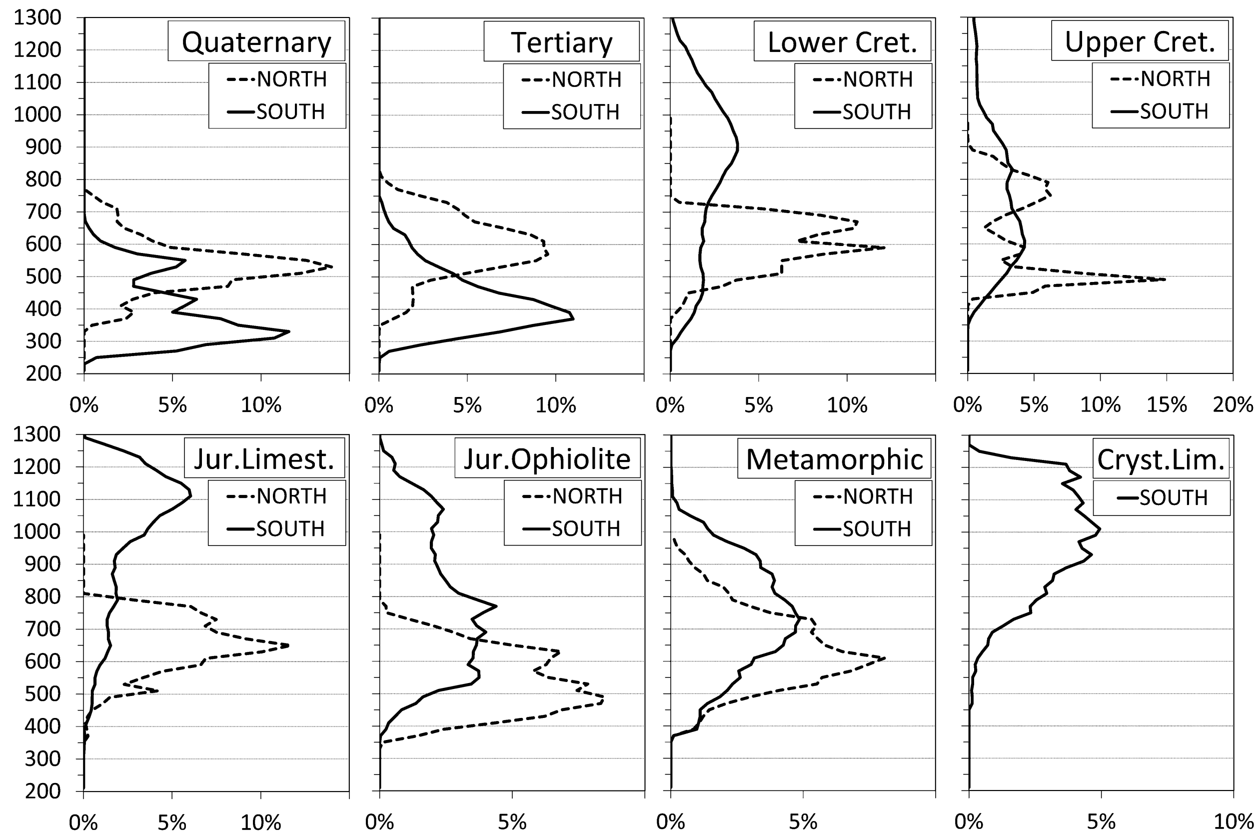

Figure 4. Elevation histograms according to rock groups.

The aspect frequencies (calculated from the smoothed DEM, see Fig. 5) show that the southern part has an unambiguous E-SE-facing main direction, but the northern part is more dispersed between ENE (60°) and SSW (190°).

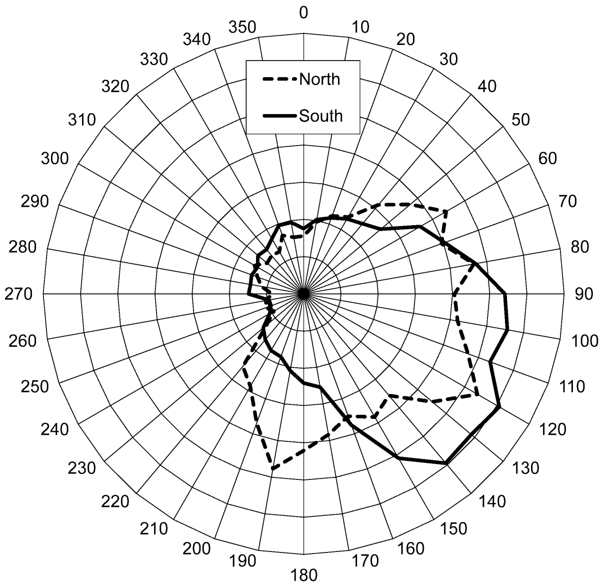

Figure 5. Aspect frequencies calculated from the mean-filtered DEM (using 1500m filter radius).

4.2. Swath analysis

Three (W)NW-(E)SE oriented swaths were analyzed in order to demonstrate general surface trends from the mountain core to the low valleys (Fig. 7B). The maximum curve (envelope surface) of the northernmost profile (Fig. 6A) shows, that the Sarmatian surface is well preserved and it is only slightly dissected. The Jurassic Limestone ridge stands out from this trend by about 120 m. The trend slope is 0.6° west of the limestone and 2.1° east of it. The minimum curve (the valley bottom of Hăşdate) is almost perfectly linear west of the limestone. In swath profiles B and C (Fig. 6) the change of surface trend is also unambiguous, but slopes are somewhat larger. The Gilăului (Metaliferi)-Bedeleu trend is 0.8-1°, whereas the Bedeleu-Mureş trend is 3.2°. The Bedeleu-Mureş trend is especially clear in swath C, it is interrupted only by limestone peaks. The linear trends suggest that vertical tectonic movements could be active where the different trends meet, that is west of the Bedeleu ridge (taking into consideration structural lines too, marked in Fig. 1B) and at the edge of the terraced Mureş Valley. The negative forms are attributed principally to differential erosion. There is a significant difference between swath B and C, namely, that the areas directly west and east of Bedeleu are much less dissected in the south than in the north. In swath C, the minimum curve has a perfectly linear section west of the limestone ridge, similarly to the northern area (in swath A).

N(NE)-S(SW) swath profiles (Fig. 7B) are perfect for detecting dissection of the main limestone ridge. In the northern part (Fig. 6D), 4 water gaps (including the Arieş) can be recognized (Fig. 7C). Beside the famous Cheile Turzii and Cheile Turului, the relatively small Borzeşti is the third. The Muntelui is possibly (but not surely) a wind gap, because the valley head is so gentle, that its formation by headward erosion is unlikely. In the southern part (Fig. 6E), 2 water gaps (Râmeţului, Intregalde) and several wind gaps are identified (Fig. 7C). The present valley bottom of the water gaps are at ~600m a.s.l., and their relative depth is also ~600m. The largest wind gaps have 150-200 m relative depth and ~1000m a.s.l. bottom level. The wind gaps are 1-2 km long valley sections with low slope angles and crossing the full width of the limestone ridge. Taking into consideration field experiences as well, it is concluded that these forms could not be created by headward erosion, but these were unambiguously formed by superposition. These wind gaps are key markers of the drainage network that started to dissect the Ciumerna-Bedeleu surface.

4.3. Analysis of the valleys

We analyzed 43 stream profiles. By using an automatic detection of knickpoints we found that 23% of knickpoints are closely related to rock boundaries, i.e. their distances to the nearest rock boundary are less than 100m. Naturally, this is a crude approximation only, since map inaccuracies may also influence this value, and in some cases, the relationship with rock boundary seems evident, but the map distance is larger than 100m. However, this value demonstrates that differential erosion is very important in the fluvial development of Trascău Mts.

Mathematical functions fitted to stream profiles show high determination coefficients (r^2>0.9, Fig. 9), however, in many cases, the type is usually not the theoratically proven logarithmic (Hack, 1973), but there is a variety of types as in the study of Carpathian rivers by Rădoane et al., (2003). They (and references therein) stated, that stream profile forms result from the action of three major controlling factors: the flow, the type of deposit over which the river flows and tectonic conditions. Here, in the Trascău Mts, we found that the linear and exponential fits are more frequent than other types. Rădoane et al., (2003) stated that linear and exponential functions are typical for rivers with coarse-grained (gravel) bed material. This is true for the streams of Trascău Mts, but local varieties can not be explained solely by this statement.

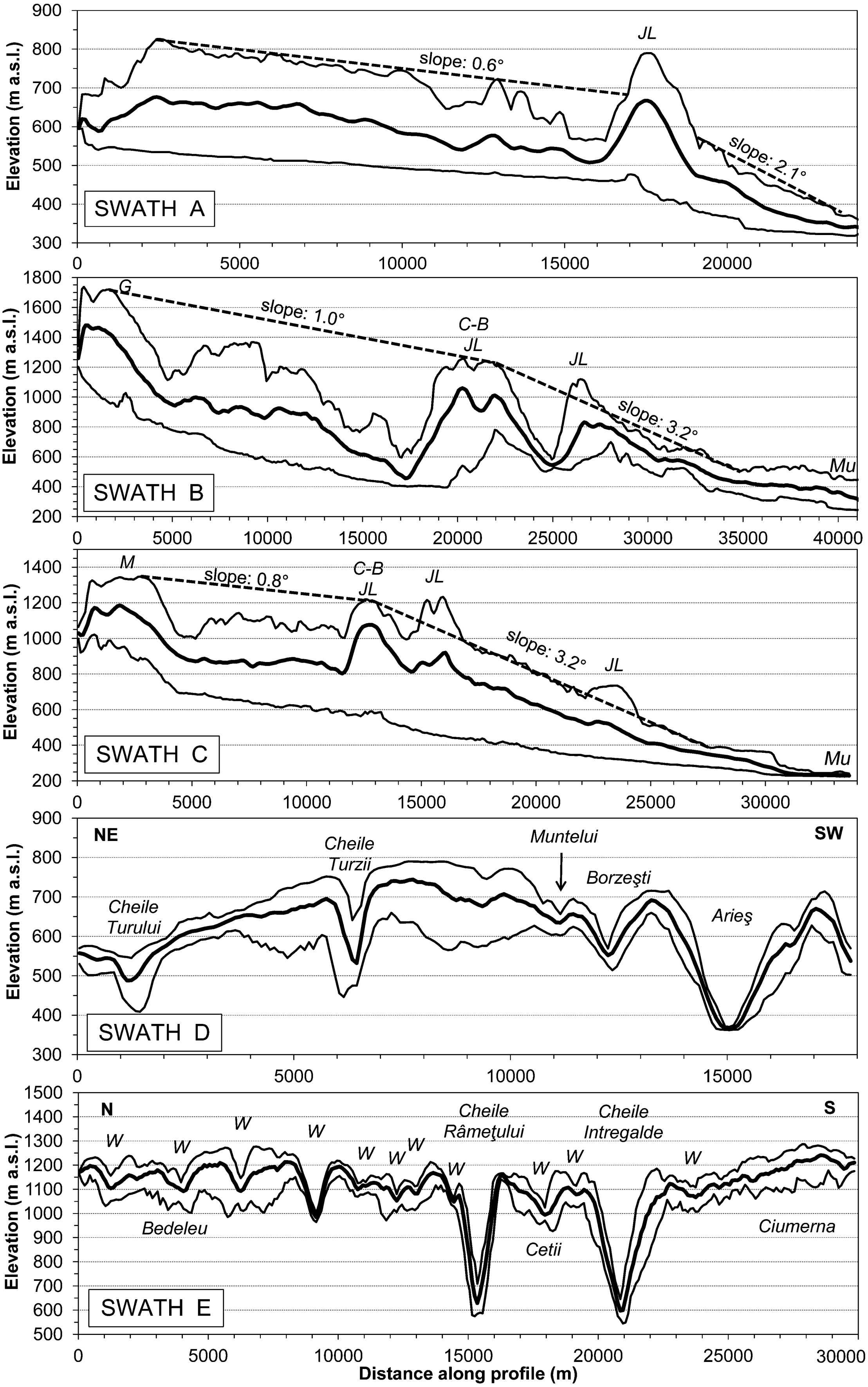

Figure 6. Swath profiles. A, B, C: W(NW)-E(SE) profiles; D, E: N(NE)-S(SW) profiles. JL: Jurassic Limestone; C-B: Ciumerna-Bedeleu; G: Gilăului; M: Metaliferi; Mu: Mureş; W: wind gap. A local maximum at Cheile Turzii in the minimum curve of swath A and small perturbations in all minimum curves are mostly SRTM-related errors.

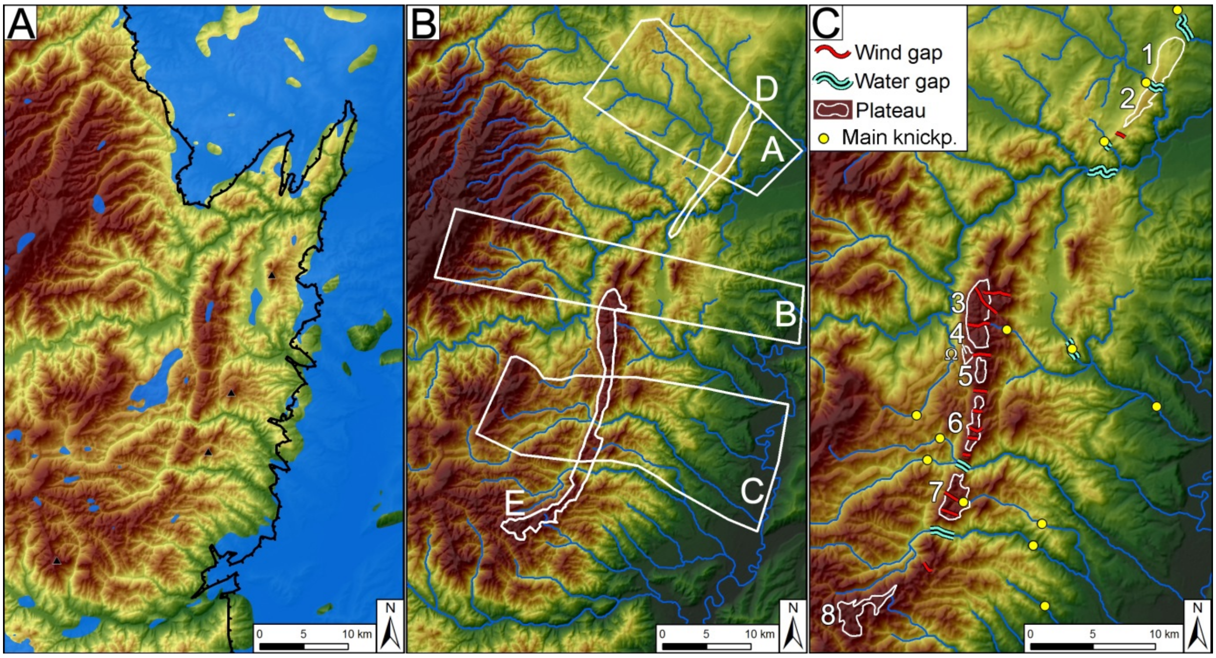

Figure 7. A: Boundary of Cenozoic rocks (black line) compared to areas where standard deviation of elevation is less than 60 m using 1500 m filter radius (blue areas); B: Location of swaths in Fig. 6; C: Water gaps, wind gaps, main knickpoints and plateaux (numbers are according to Table 1), Ω marks Huda Lui Papară cave.

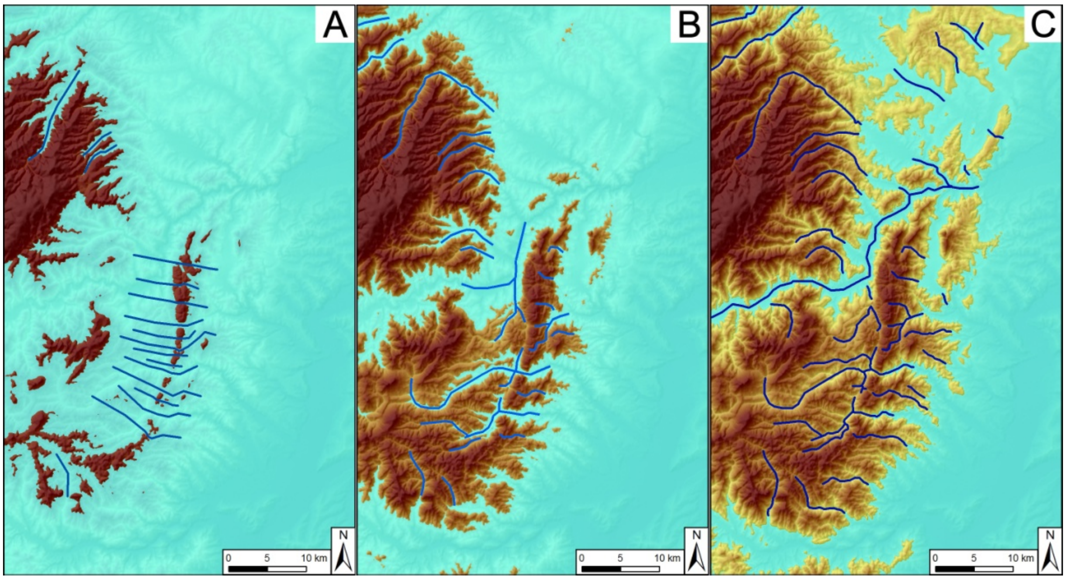

Figure 8. DEM filled up to certain levels and stages of valley development. Fillup levels are 1100m a.s.l. (A), 800m a.s.l. (B) and 600m a.s.l. (C), respectively.

Based on the general shape of stream profiles and neglecting smaller knickpoints, we found three types.

1. Simple, linear (Fig. 9A). It is typical for smaller streams, and the only example among longer streams is Galda (except its upstream end). It remains a question why it is different from other streams crossing the main limestone ridge.
2. Simple concave (Fig. 9B). It is also typical for smaller streams, especially west of the main ridge, where the Râmeţ-Ponor surface is found. An important, long river belonging to this group is Iara (but not a perfect example, because it can be better modelled by a compound exponential and a linear segment). This type is thought of as the most regular one (Hack, 1973). It can be formed where disturbing effects (rock differences, tectonics) are neglectable and the adaptation time is long enough.
3. Double concave (Fig. 9C, D). It has two subtypes:
   a. The main knickpoint (Fig. 7C), i.e. the boundary between the main segments is connected to the section crossing the

limestone ridge. Low slope and an almost linear profile section is typical for a certain distance upstream of the watergap. Examples are Racilor, Hăşdate, Aiudul.

b. The main knickpoint (Fig. 7C) can not be attributed to rock boundary. In these cases, the knickpoint is likely to be a tectonically induced, receding knickpoint. Typical examples are found at the southeastern part (e.g. Cetea, Neau, Ighiu).

Other short-length, typical disturbances are also observed on stream profiles. First, streams starting from wind gaps have a convex segment at their beginnings (e.g. Bedeleu, Cetea, Fig. 9E, D). Second, streams crossing the ophiolite terrains have an undulating character (e.g. Muntelui, , Fig. 9F). It is the result of the heterogeneity of ophiolite rocks, which is supported by field evidence, too.

The cross-sectional size of a valley is basically a function of discharge, age and rock type. Discharge, in turn, is a function of drainage area. We compared the Inzelului and Geoagiului valley cross-sections east of the main ridge. Both valleys are in Upper Cretaceous rocks, the valley dimensions are similar, therefore the corresponding drainage areas should be also similar. However, we found that at present, Geoagiului drainage area ($144.8 km^2$) is 3.75 times larger than that of Inzelului ($38.6 km^2$). Therefore, it is concluded that the Inzelului drainage area had to be also larger during a significant part of valley formation. So this fact supports the idea that wind gaps once hydrographically connected the areas west and east of the main ridge.

In order to approximate stages of valley development in uplifting mountains, it is a useful tool to fill up the DEM to certain levels (Fig. 8). These may be tought of as base levels, and higher terrains as hills or islands standing up from the surrounding lowlands or seas. Of course, it is a crude approximation, since tectonic movements and differential erosion may have also caused significant differences, however it may help to outline certain features of the palaeo-drainage.

The first fillup level is at 1100m a.s.l. This image (Fig. 8A) shows the remnants of the Carpathian pediplain surface. Remnants of the valleys are the present-day wind gaps in the Ciumerna-Bedeleu ridge, and the uppermost segment of Iara and some other smaller streams in the Gilăului Mts. Altogether, a radial drainage pattern can be hypothesized for this stage.

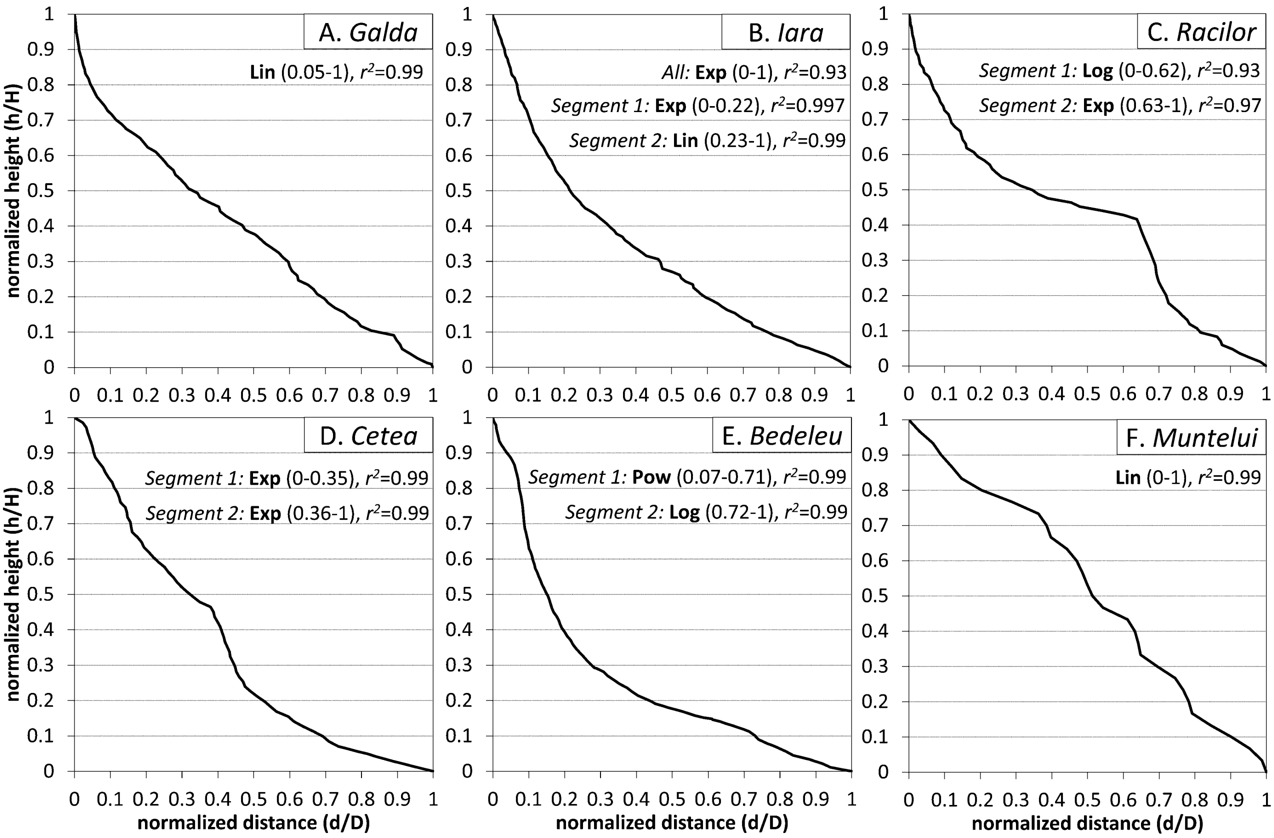

Figure 9. Selected normalized stream profiles with type and r^2 of best-fit functions (fitting segment limits are written in brackets). A. Simple, linear; B. Simple, concave; C. Double concave with main knickpoint at the limestone ridge; D. Double concave with main knickpoint on homogeneous lithology; E. Profile with convex upstream segment; F. Undulating profile.

The second fillup level is at 800m a.s.l. (Fig. 8B), which is about the present-day Râmeţ-Ponor surface level. The main ridge forms an obstacle and drainage is diverted, therefore a trellis-like drainage pattern is evolving. Several former water gaps became wind gaps by this stage. In the Gilăului Mts, there is an abrupt bend in the course of Iara valley and also in the neighbouring smaller streams. This bending is found on a lithologically homogeneous area. The northern part of the study area is almost totally below this level, and only a minor portion built up of Sarmatian rocks is standing out from this level. Based on this, that stage can be attributed to late Sarmatian.

The third fillup level is at 600m a.s.l. (Fig. 8C). We can observe the consolidation of the previous drainage pattern, but many features are fixed only during this stage. Superposition of several present-day water gaps (Cheile Turzii, Cheile Vălişoarei, Arieş at Buru, lower Iara valley) begins in this stage. In the northern part, there are still large indeterminate areas, the incision of Racilor stream has not yet reached the limestone. It is possible that Hăşdate drained a part of the Gilăului Mts. And in general, the whole eastern sector of the study area is under this level. The final step is the present (Fig. 1), when the erosion base level is at 230m a.s.l. at the Mureş valley and at 330m a.s.l. at the lower Arieş valley.

4.4. Karst landforms

On elevated karst terrains, exokarst and connected endokarst landforms substitute the drainage network. Further on, karstification could play an important role in the formation of the actual river network, too. This is why it is important to study karst landforms as well in order to better understand drainage network evolution in Trascău Mts. Among exokarst landforms, we studied the dolines. Previous geomorphometric research demonstrated that most dolines develop on low-slope limestone surfaces (e.g. Orndorff et al., 2000; Telbisz et al., 2007). Thus, we delineated areas that are suitable for doline formation as limestone terrains with less than 12° slope angle (Fig. 7C). Doline densities were calculated for these areas (Table 1). Field GPS measurements show that real doline density is ~2-fold of the value calculated based on the 1:25 000 scale topographic map, because smaller forms are not marked in the map. Furthermore, this results that mean doline area is smaller when calculated from GPS data. In general, it is stated that doline density is relatively poor with respect to similar karst regions (e.g. Telbisz et al., 2007; Telbisz & Ádám, 2011), and there is a remarkable difference between the northern and the southern parts. The real doline density is 6.6-12.7 km^{-2} for the southern plateaux, while it is only 0-0.6 km^{-2} for the northern part. Although this low density is usually attributed to the small width of the northern plateaux, it does not hold true, since similar width in the southern parts are linked with higher densities.

Based on the high length/width ratio of the limestone terrain and the long contact with the surrounding non-karstic rocks we could name the main ridge a contact karst and even a stripe karst (cf. Lauritzen, 2001; Ćalić, 2001). However, while stripe karsts are usually rich in exokarst and endokarst features due to allogenic rivers, here, in the Trascău Mts, real contact karst phenomena are relatively rare. There is one nice example of a through cave (Huda Lui Papară, Fig. 7C) with a large-size ponor, but it crosses only a western tip of the limestone terrain and not the main ridge. Otherwise, there are no ponors at the western, upstream part of the main ridge and there are no through caves below the limestone ridge. One potential reason is the WNW dip direction of limestone strata and the steep dip angles that did not favour the formation of west-east oriented through-caves.

Table 1. Morphometric data of dolines

Plateau	Area (km^2)	Width (m)	Doline density, map (km^{-2})	Doline density, GPS (km^{-2})	Mean doline area, map (m^2)	Mean doline area, GPS (m^2)
1.Turda, North	5.1	1750	0.0	0.6	n.d.	1758
2.Turda, South	2.1	800	0.0	0.0	n.d.	n.d.
3.Bedeleu	8.1	1720	3.2	6.6	7644	4537
4.Secului, NW	0.8	810	5.1	n.d.	3414	n.d.
5.Secului, Main	1.2	800	4.0	n.d.	3581	n.d.
6. Geamânului-Trascâului	2.9	870	5.6	n.d.	3864	n.d.
7.Cetii	5.2	1520	5.8	11.8	2982	2344
8.Ciumerna	5.6	1200	2.9	12.7	1803	803

5. DISCUSSIONS

An important fact, the ~200m higher position of Cenozoic sediments in the northern part may be caused by lower denudation rate or by relative uplift. Among these factors, the lower denudation rate can be explained by the *blocking* effect of the limestone ridge, since headward erosion of streams flowing to the subsiding Corneşti basin (lower Arieş valley) is effectively blocked by the limestone as it was demonstrated by the stream profiles and swath analysis. These results are in agreement with the opinion of Cocean (1988). Second, the Iara valley drains all valleys from Gilăului Mts flowing towards NE that limits the drainage area belonging to the northern part, therefore the actual fluvial erosion is also limited in the northern part.

The Iara bend (which is marked by a knickpoint in the stream profile) and the strong planform linearity of the downstream Iara valley suggest structural control (an idea that was first presumed by Kerekes, 1921, see Fig. 1B). Furthermore, this line is recognizable to the northwest and to southeast as well in smaller stream sections and in SW-facing slopes that further supports its structural origin. The main axis of Hăşdate stream (including Cheile Turzii) is also linear and parallel with Iara, i.e. it may be also tectonically pre-determined. Another argument for the structural control is the low slope of the envelope surface (Fig. 6A) that, in itself, should have led to the formation of a more dendritic drainage pattern instead of the present (sub)-parallel one. Thus, trying to interpret these structural lines and taking into consideration the west-east compression in late Miocene times, these lines may be attributed to antithetic faults in a west-east compressional stress field. However, the geologic maps (Giuşcă & Bleahu, 1967; Kounov & Schmid, 2012) does not indicate structural lines here and further study is necessary to clearly resolve this problem.

The good agreement of low slope (or low standard deviation of elevation) area and the superficial extension of Cenozoic rocks can be explained by two hypothesis. First, Tertiary and Quaternary sediments have lower rock resistance, therefore the resulted slope angles are lower. Second, valleys incised in Mesozoic and older rocks had longer duration of fluvial evolution.

The aforementioned facts about karst landforms (e.g. the relatively low doline density, especially in the northern part and the lack of ponors) suggest that dolines on the limestone plateaux were formed mostly by autogenic karst processes. This way, differences in doline density between the northern and southern parts may be explained by the different time duration of subaerial karstification.

Futhermore, the lack of ponors and through-caves makes it likely that water gaps and wind gaps were formed predominantly by superposition. However, it does not exclude that for certain periods subterranean pathways could develop below the superposed valleys and these pathways could later collapse. Briefly, valley formation by cave collapse and valley formation by superposition are not controversial processes (for further explanation see Hevesi, 2000). In the Trascău Mts, the driving force was superposition and cave collapse could be occasional, the only unambiguous evidence being the rock portal in Cheile Râmeţului. Since uplift was almost continuous during the drainage evolution of the study area (see Section 2), antecedence was also important in the formation of almost all gorges.

Taking into consideration all of the above facts, it is concluded that tectonic movements and high rock resistance of limestone were the main reasons for the diversion of rivers (as outlined in the previous section), meanwhile intense karstification and underground capture of surface streams (as emphasized by Cocean, 1988) could play only a subordinate role.

6. CONCLUSIONS

In most cases, extended limestone plateaux (e.g. Canin Mts in Slovenia-Italy, Telbisz et al., 2011) have an unambiguous signal in the elevation distribution. However, in case of the Trascău Mts, histogram analysis demonstrated that the Ciumerna-Bedeleu surface is hardly recognizable in the histogram due to its limited extent. On the other hand, elevation and slope histograms proved to be very useful in the exploration of differential erosion (and potentially tectonic) phenomena. Aspect frequency analysis demonstrated that the southern part has a predominantly E-SE consequent flow direction, which constrains the direction of larger valleys. On the other hand, subsequent valleys are rather adjusted to the N-S orientation of rock stripes and in many cases have simple stream profiles.

Swath analysis demonstrated that there is a characteristic change in the slope of the envelope surface at the Ciumerna-Bedeleu surface and at the edge of Mureş valley. These changes are attributed to tectonic movements. N(NE)-S(SW) swaths profiles helped the recognition and quantification of water gaps and wind gaps. Especially, these latter forms were neglected in previous literature, in spite of the fact, that these are very important remnant landforms of the post-Cretaceous drainage network.

The most typical stream profiles of the study area can be modelled by exponential and linear functions and several stream profile types (simple-linear, simple-concave, double-concave) could be identified. The large number of bedrock-related knickpoints also highlights the importance of differential erosion in the landform development of Trascău Mts. The denudation blocking effect of the main limestone ridge was also clearly detectable in the stream profiles.

The evolution of drainage network was approximated by a DEM fillup technique. The original Post-Cretaceous radial pattern evolved to a trellis pattern (due to differential erosion and probably to differential uplift). It is concluded that superposition (with antecedence) played the most important role in the formation of water gaps and cave collapse had limited significance only.

Practically all analysis highlighted the difference between the northern and southern parts. These parts had similar landform evolution, but are at different stages. The Post-Sarmatian evolution of the northern part copies the Post-Cretaceous evolution of the southern part. This explains why larger exokarst landforms (dolines) and wind gaps are almost missing in the northern part.

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