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Tectogeomorphologic Geometamorphoses in Northwestern Romania – the Sylvania Mountains

Ion MAC

Key words: morphotectonogenesis, basins and catenas, dislocated Hercynian orogen, Sylvania Mountains, palimpsestic relief.

Abstract. Apparently uniform and homogenous, the territory between the Rodna Mountains (Eastern Carpathians) and the Central Apuseni Mountains, namely the Plopiș and Meseș Mountains, is described by the presence of some old tectostructures and paleo-orogens under sedimentary layers of tertiary sandstones. Three major statements regarding this territory were advanced in the geography-geology field:

- a. Linking unit between the Apuseni Mountains and the Eastern Carpathians, called by V. Mihailescu (1921, 1930, 1935-1936) “the Intra-Carpathian yoke”;
- b. Carpathian subunit, linked to the Apuseni Mountains (Coteț, Martiniuc, 1960);
- c. A complex structural block developed as a result of Sylvania’s lifting up, with a distinct evolution (Mutihac, 1992).

Within another conception, the mentioned territory could be attributed to a mountainous system, linked, as regarding their genesis and evolution in time, to the Eastern Carpathian through the Rodnei Mountains and to the Apuseni Mountains through Gilău and Bihor Mountains down to the Banat Mountains. This Hercynian orogen was fragmented, sunk and then involved into the tectostructural Tertiary movements. The posttectonic movements fragmentized the respective system and determined its affiliation to the Transylvanian Depression rigid craton, considered a remaining of Gondwana. The Sylvania Mountains, with a geomorphologic landscape resembling to that of the European basins and catenas, is the result of this geologo-geomorphologic evolution and changes. The initial Hercynian tectogene was intensely modeled in subaerial regime determining a palimpsestic morphology unique within the Romanian territory.

1. General Considerations

Apparently uniform and homogenous, the territory between the Rodna Mountains and the Central block of Apuseni Mountains, namely the Plopiș and Meseș Mountains, “screens” some old tectostructures and paleoreliefs by sedimentary layers of Tertiary sandstones.

The most recent geologic researches on the mentioned area, focuses with predilection on the crystalline formations and on the sedimentary layers (for example Săndulescu, 1984, Balintoni, 1994, 1996), led to the conclusion that, excepting the Gilău Mountains, the crystalline formations of the “Someș series” (belonging to the Bihor Unit – the Bihor Autochthonous) afflorates on the ridge of the Remeti graben within the Rez and Meseș Mountains (Plopiș) and within the Faget and Țicău crystalline isles. As regarding the Preluca Mountains, the opinions are contradictory, according to some of them they belong to the Baia de Arieș series (Roșu, 1983), while others associate them to the Rebra series being considered as a component of another unit, different from the Bihor one.

Regardless of the asserted opinions, it is obvious that the Preluca Massif keeps orographically and morphologically a landscape that resembles to that of the above mentions crystalline isles.

The crystalline formations are appreciated to be Precambrian or possibly Carboniferous. On this crystalline layer, Postsenonian Mesozoic formations and locally Permian ones are sporadically found.

The second stratigraphic sequence belongs to the Superior Triassic and the Low Cretaceous. A distinct prealpine paleogeographic situation was thus identified, with obvious implications in the actual morphology.

The Alpine morphologic and tectogene deformations induced structural and geomorphologic readjustments, reactivation of old faults and dislocation of the existing deposits, especially of the Tertiary formations.

When analyzing the evolution of the geologic-geomorphologic events, the following phases could be distinguished:

- a) The Danian-Paleocene phase developed at the end of the Superior Cretaceous and continued in

the Paleocene, when the marine transgression from the Transylvanian basin took place, contributing to the creation of a communication way towards West through the Șimleu Basin. The correspondent formations are to be found in the Jibou-Benesat-Ulmeni area where they form the Danian-Pliocene layers with their well-known series (Joja, 1956);

- b) The basin phase (Badenian-Sarmatian-Panonian) characterized by transgressions (for example the Badenian or Panonian transgression) and regressions (Upper Sarmatian, Pontian) which determined an uniform morphology, made of ridges, insular massifs and basins that induced a differentiated sedimentation and modeling.
- c) The subaerial modeling phase (morphosculptural), developed along the Upper Tertiary and Quaternary, described by processes of selective erosion, large denudation or fluvial modeling that resulted in the appearance of extended piedmontan layers, fluvial terraces and altitudinally ordered glacises. The region's actual geomorphic physiognomy is an outcome of the processes that took place at the end of this phase.

2. Geologic and geomorphologic References

Along the Precambrian and Paleozoic eras, the analyzed region belonged to a large geosynclinal area.

The Hercynian orogenesis (Middle Carboniferous) determined the formation of a mountainous system and the transformation of the studied territory into a rigid block that under the action of a subaerial modeling would have become a *pediplena* (Pop, 1964).

The Alpine orogenesis subjects the Hercynian morphotectonic edifice to a lateral compression, transforming it into an "isle" area – horst, ridges and local basins developed on the sinking microblocks. The end of the Alpine orogenesis (the Attic, Rhodanic and Valachian movements) also induces plicative and fractural deformations of the Neogene deposits in the areas of marine and lacustrine accumulation, the grabens that separate the "catenas" and the horsts.

The significant lifting up at the end of the Alpine orogenesis (the Valachian movements) endows the old structures on the alignment Meseș - Dealu Mare - Preluca with a remarkable elevation that is physiographically expressed through a

mountainous configuration. As a result, erosion increases, contributing to the removing of the Neogene formations and to the formation of the leveled surfaces at 300-350 m and at 550-600 m. The hydrographic network that models the contact depressions developed in the subsidence areas corresponding to the grabens: Lower Someș Depression, Jibou-Ulmeni Depression, Agrij Depression and the internal depressions of Simleu, Zalău și Sălaj.

3. Typology and Functions

The following opinions were advanced in the geologic and geomorphologic literature in relation to the genesis and spatial function of this territory:

- a) The hidden mountains of the Northern Transylvania (Szadecky, 1913, Berindei, Mac, 1980);
- b) Linking unit between the Apuseni Mountains and the Eastern Carpathians, called by V. Mihăilescu (1921, 1930, 1935-1936) "the Intra-Carpathian yoke";
- c) A complex structural block developed as a result of Sylvania's lifting up, with a distinct evolution (Mutihac, 1992);
- d) Pericarpethian unit belonging to the Banat-Criș hills and depressions (Pop, 2005);
- e) Excepting the Meseș and Plopiș Mts., considered as a prolongation of the Apuseni Mts. orogen, many researchers consider the rest of the analyzed territory as a hilly (piedmontan) unit and name it the Silvano-Someșan Hills (Mihăilescu, 1936, 1966, Mac, 1991).

Other assertions are to be found in the specific literature, but we consider them formal and strictly contextual (Pop, 2000).

4. New Approaches and Interpretations

As a result of the field survey, the following elements were identified:

- a) The Meseș Mts, Dealul Mare Hill, Țicău Hill and Preluca Hill are described by Paleogene formations that were strongly disturbed tectonically, vertical strata being identified in some places. Thus, a large monocline developed towards the Transylvanian Plateau and a composite cuesta front towards Sylvania. The following question arises from this situation: how and when such disturbing phenomena

could happen? The answer for when is quite obvious, in Postoligocen;

- b) All the ridges (Meseş, Plopiş, Codru) and the crystalline isles are asymmetric: a steep slope to the West or Southwest and a gentler one to the East and Northeast. This clinotropy supposed a pushing impulse/effort from West and Northeast and a rigid barrier in the East, from the Someşan Plateau;
- c) The relief on the crystalline formations is ordered vertically in general on two leveled surfaces, while on the adjacent sedimentary another one has developed, at 300-350 m;
- d) The main hydrographic network has a general orientation from South to North (the Someş,

Crasna, Barcău, Sălaj rivers), that is a longitudinal one, while the secondary net is perpendicular, with evident transversal sectors in the case of many tributary rivers (Cerna, Oarţa, Sărata etc.)

- e) Beneath the ridges' and interfluves' level, the denudation process gave birth to basal or slope glacises extremely well represented on the edge of the external and internal depressions.

The geologic and geophysical data reveals the following facts (Fig. 1):

- a) The crystalline formations form ridges and hummocks ordered longitudinally along a North-South direction;

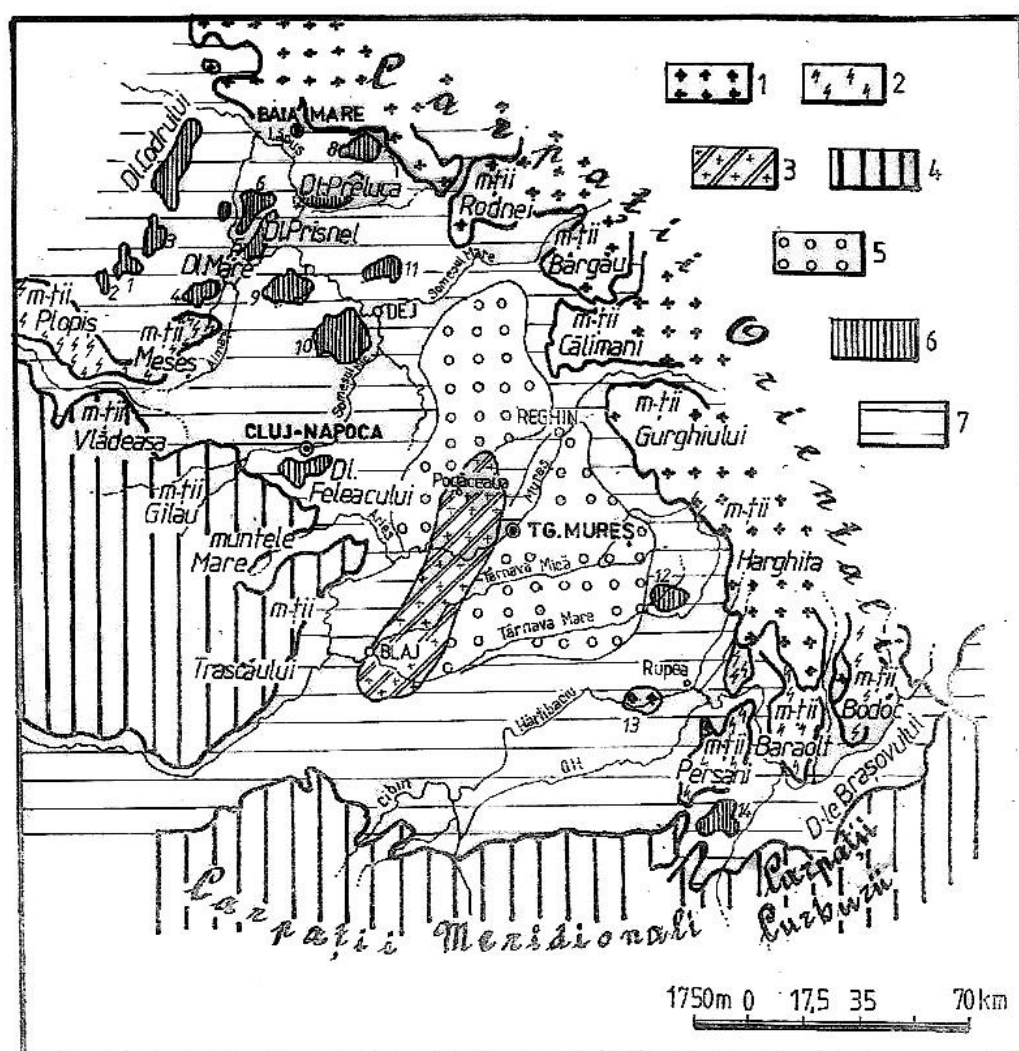


Fig. 1. Tectostructural framework of the Transylvanian Depression and of the Sylvania Mts.

1. Magmato-volcanic structures – Eastern Carpathians and Rupea Isle; 2. Crystalline and sedimentary structures from the marginal orogens; 3. Areas of neotectonic lifting up and Miocene rocks; 4. Complex Carpathian orogens (Apuseni Mts., Southern Carpathians); 5. Mio-Pliocene formations; 6. Structural hummocks and ridges (1 – 14); 7. Transylvanian Depression and its connection units.

- b) A series of tectonic dislocations oriented almost from East to West and fragmenting the crystalline were identified (for example the Moigrad fault);
- c) Between the line of ridges and hummocks, sedimentation areas developed, especially with Tertiary (Neogene) formations (Fig. 2);
- d) The sedimentary formations are folded or monoclinal. An example in this direction is the Cuceu-Benesat-Ulmeni, the Aghireş-Panic, the Şamşu-Hodod and Mirşid-Doba Anticlines;
- e) A general subsidence of the whole tectostructural ensemble towards West-Northwest is to be noticed. The fault (the deep fracture) Carei-Oradea marks the transition from the Silvania tectonic unit to the Pannonian one;
- f) Two major phases describe the evolution of this region: the Predanian and the Alpine one.

5. Final considerations

Associating the opinions of some researchers (Coteţ, and Martiniuc, 1960, Mutihac, 1990) with the geologo-geomorphologic and cartographic evidences, as well as with the available information on the region's global tectonics and the field

surveys, the following tectonic and geomorphologic considerations could be asserted:

1. The studied geologic and geomorphologic unit appears as a distinct territorial entity in Northwestern Romania, that we named as the Silvania Block Mountains (or the Silvania Mts.). They stretch from the Lăpuş Depression (Posea, 1962) in the East to the Crişul Repede Valley in the West.
2. Major faults and fractures mark the extremities of these block mountains: the Olt-Huedin-Crişul Repede-Oradea fracture to Southwest, the Lower Someş fracture to Northeast, the Carei-Oradea fracture to the West and the Parameseşan fracture to the East.
3. The territory is described by the features that characterize the block mountains, benefitting from an architecture of horsts and grabens. The horsts appear as mountains (Meseş, Plopiş, Preluca, Țicău, Prisnel, Codru) or hummocks (Simleu, Cosei, Dealu Mare, Dealu Chicera etc.). The grabens are strongly sedimentated and generally modeled as longitudinal valleys, some of them with fluvial terraces (Crasna, Zalău, Sălaj, Someş).
4. The rivers leave the block mountains through gorges (Marca, Porți, Țicău).

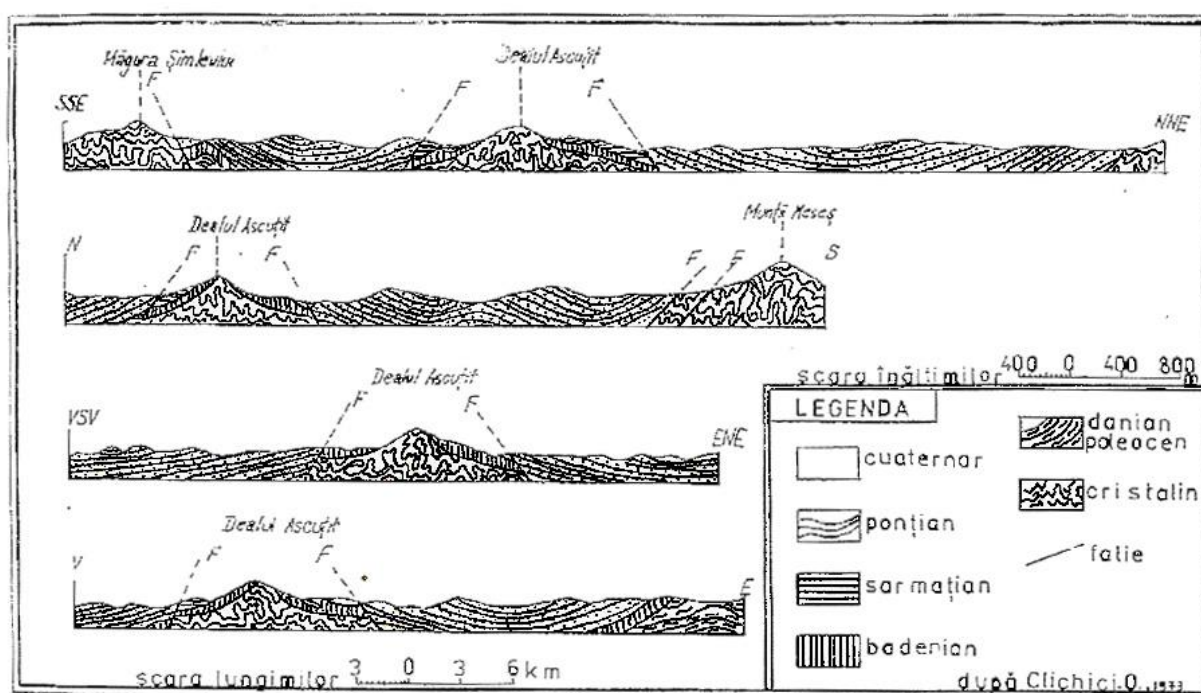


Fig. 2. Transversal sections in the Șimleu Depression and in its marginal sectors, on various directions

5. The genesis of this unit should be attributed to the tectonic and orogenic processes that took place in the Proterozoic and the Carboniferous, especially to the Hercynian Orogenesis, that acted between two central blocks: the Transylvanian central block, interpreted today as a remaining of Gondwana in its advance to the East-European Plate and the Pannonian block, a small traveling plate under the impulse of the Atlantic Rift. The first crustal block, Transylvania, stopped the developing of Hercynides towards South, because of its rigidity due to the incorporated old and new orogens. This fact explains the altitude of the ridges Meseş, Prisnel-Țicău-Preluca-Rodna and the strong distortion of the Paleogene deposits in the southern part of the latter mentioned unit. These formations were crushed towards the Transylvanian craton under the impulse of the orogenetic movements.
6. The system of the Sylvania Block Mountains includes several subunits:
 - a) The marginal mountainous subunit composed, at its turn, of some subunits: 1.1. Plopiș subunit; 1.2. Meseş s.; 1.3. Prisnel-Țicău s.; 1.4. Preluca s.; 1.5. Codru s;
 - b) The intra-mountainous depressions subunit: 2.1. Simleu Depression; 2.2. Zalău Depression; 2.3. Someș Depression (Jibou-Ulmeni);
 - c) The hummocks, hills and piedmontan depressions subunit : 3.1. Simleu Hummock; 3.2. Coseu Hummock; 3.3. Chicera Hummock (to the North of Jibou);
 - d) The valley corridors and erosion saddles subunit : 4.1. Hurezu Mare-Bicaz-Oarța Passage; 4.2. Bogdand-Supur Passage; 4.3. Cehu Silvaniei Passage; Borla-Sărmășag-Supur Passage.

The increased morphologic dividing constitutes the reflex of the tectonic and hydrographic fragmentation and expression of the differentiated modeling within the Sylvania Mountains. Together with the morphologic contrasts imposed by structure and rock, some gentle physiographic elements also appear as a result of the sedimentary deposits modeling in the local sedimentation basins.

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Utilization of Geoinformatic Methods in the Morphometric Analysis Case Study on a Mesa from Hungary: the Somló Hill*

Titusz, BUGYA, István Péter, KOVÁCS

Key words: dry valley, morphometry, contour line section index, GIS, Somló Hill

Abstract. Somló situated on the Little Hungarian Plain (*Kisalföld*), represents an ideal territory for geomorphologic and morphometric analysis. In the course of the research, the aim was to work out an unambiguous descriptive method applicable to the two types of derasional valleys of the sample territory. Therefore, the already existing traditional geomorphologic results, as well as statistical and geoinformatic techniques were utilized.

In the first stage of the analysis, the characteristic features of each section of the contour lines—crossing the dry valley—were examined. The average contour line section index (CLSI) and the standard deviation of each young and old valley was defined, according to which the valleys belonging to these two groups as well as the morphometrically different valleys could be separated from each other. The extent and location of change observable in the cross-section of valleys was also described.

In the second stage of the research, the characteristic inclination of the valleys was analyzed. According to the maximum values of the normal distribution projected on the slopes of the valleys, it was also possible to distinguish the old and young derasional valleys. The valleys with characteristic values different from that of the above mentioned two groups were revealed as well.

Introduction and Aims

Somló Hill (432 m) is situated on the Little Hungarian Plain, in the western part of Transdanubia, in Hungary (Fig. 1). Its major building rocks are easily eroding sediments of Lake

Pannon deposited in the Central Paratethys. The lake retreated from the research area approximately 9–9,5 million years ago (Magyar et al., 1999). The deposits consisting of clays and sands were covered by hard, resistant basalt, which protected them from the effects of the exogenic processes.

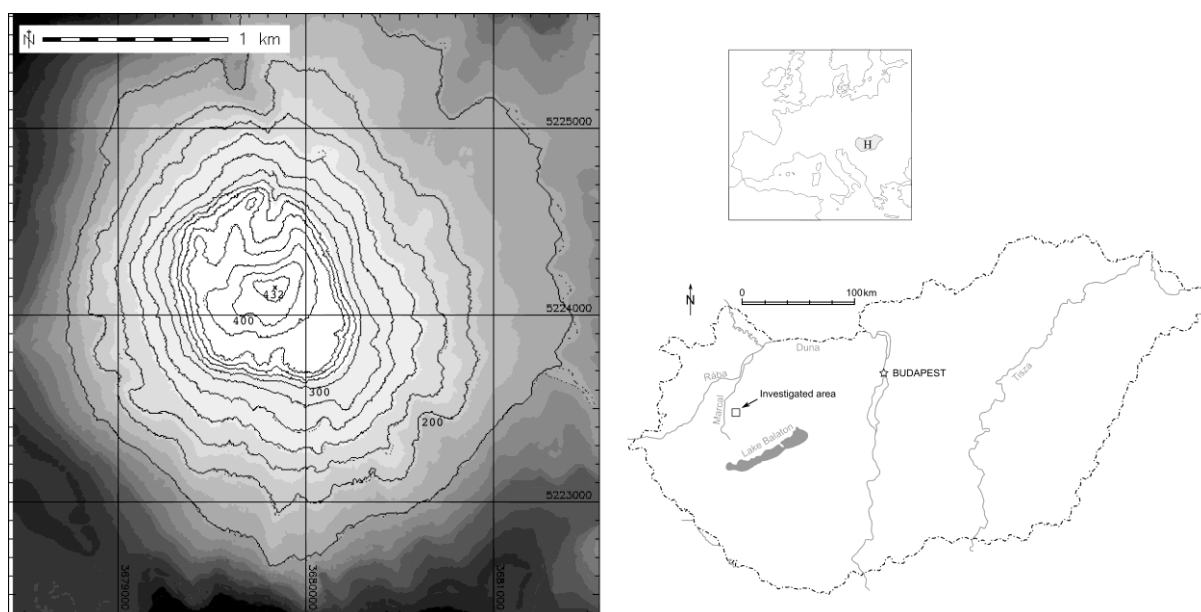


Fig. 1. Location of the study area. (UTM coordinates)

* Paper presented at the “IAG Regional Conference on Geomorphology: Landslides, Floods and Global Environmental Change in Mountain Regions”, September 15-25, 2008, Braşov, Romania

The duration of the Somló's volcanic activity is still doubtful. The researches based on absolute dating (Balogh et al., 1982, 1987; Borsy et al., 1986; Pécskay et al., 1995; Wijbrans et al., 2007) define the period lasting for 3,5 million years. The relative chronological interpretations (Varga et al., 2004, Schweitzer, 1997) and the latest geomorphologic observations (Kovács, 2008) calculate with an older volcanic activity dating back minimum 5 million years. These latter examinations and results were utilized in this research.

After the basalt volcanic activity had finished, the hill grew high above its environment, and on it developed two generations of derasional valleys. These valleys represent a special group of dry valleys. They were formed by sheet erosion and mass movements, under a Pleistocene periglacial climate. This study utilises this term originating from Pécsi (1964/a, b, 1997) as well.

The young derasional valleys developed on the deposits of Lake Pannon, resulting from Pleistocene periglacial processes. The old derasional valleys were deflation hollows incised in the basalt mesa (Kovács, 2008), and were formed after their dissection under periglacial climate alike (Fig. 2).

Morphometric parameters and evolution of derasional valleys were particularly investigated in the last 50 years in Hungary. Cross-sections, slope maps and geomorphologic sketches were made (Kéz, 1956, Marosi, 1965, Szilárd, 1965, Pécsi, 1962, 1997), though their detailed description and analytic morphometric examination is still not completed. However derasional valleys play an important part in flash foods (Czigány, et al. 2008). They cover more than 60% of Hungary's surface (Pécsi, 1997).

In the course of the research, the authors' aim was to work out an unambiguous descriptive method applicable to the two types of the dry valleys of the sample territory, using traditional geomorphologic results (geomorphologic map), statistical and geoinformational techniques.

Methods

In this work the 6.0 version of the Grass GIS software was utilized, which was run in a Debian Linux environment. The contour lines of the topographic map of the research area to a scale of 1 : 10 000 were digitalized on-screen, using the *v.digit* routine of Grass 6.0. The resulting vector lines were rasterized using the *v.to.rast* routine.

Grass 6.0 creates a raster elevation map from a rasterized contour map using the *r.surf.contour* routine. To determine the elevation of a point on a contour map, we could interpolate its value from those of the two nearest contour lines (uphill and downhill). The *r.surf.contour* routine linearly interpolates between contour lines (Ehlschlaeger, 2007). On the topographic maps used contour lines are indicated every 2.5 meter, therefore the theoretical breaking down of the resulting elevation map includes 5x5 meters. The geomorphologic map of the Somló was constructed based on the above mentioned topographic map, thus its digital version could be compared with the elevation map of the hill. The young and the old derasional valleys represented on the geomorphologic map were distinguished and grouped using the *r.reclass* routine.

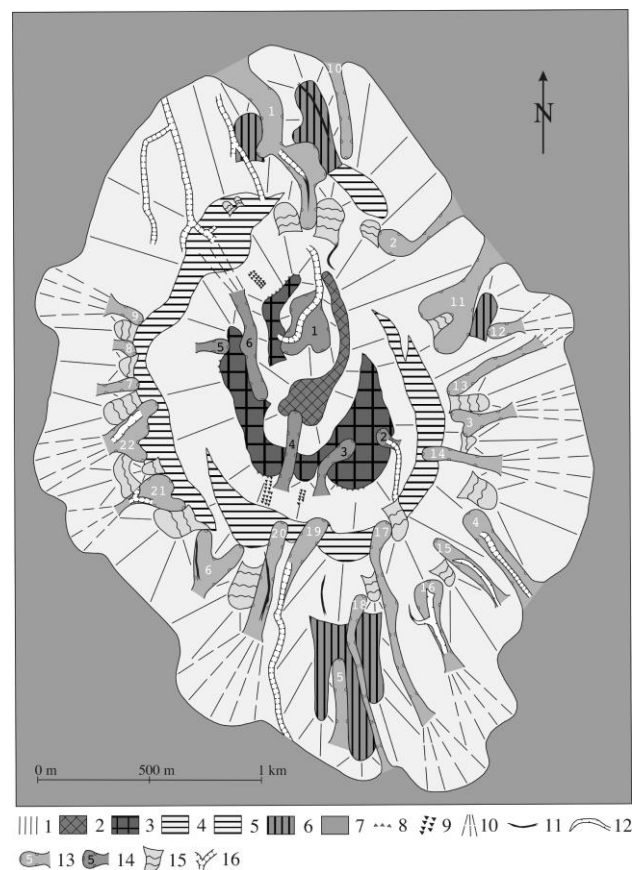


Fig. 2. Geomorphological map of Somló Hill and Id. numbers of the derasional valleys

1 = slopes undistinguished; 2 = top region of the mesa;
3 = mesa; 4 = higher level of pediment; 5 = lower level of pediment; 6 = interfluvium; 7 = alluvium; 8 = columnar basalt;
9 = debris flow; 10 = debris cone; 11 = gully; 12 = ravine;
13 = young derasional valley and Id. number; 14 = old derasional valley and Id. number; 15 = landslide; 16 = defile
(After: Kovács 2008)

In the first stage of analysis, the characteristic features of each section of the contour lines—crossing the derasion valley—were examined (Fig. 3). The geomorphologic map was completed by the projection of the rasterized map of contour lines, and the data were measured on this version. Subsequent to the input and run of the “*d.measure - m >>*” routine, the length of the contour lines in the

valleys was measured with the mouse, and listed into a file. The second step was to define the length of the straight line between the incoming and outgoing point of the contour lines. These two kind of measurements were repeated in every case of contour lines crossing a derasional valley. Beside the length of the measured sections, the elevation values of each contour line were also indicated.

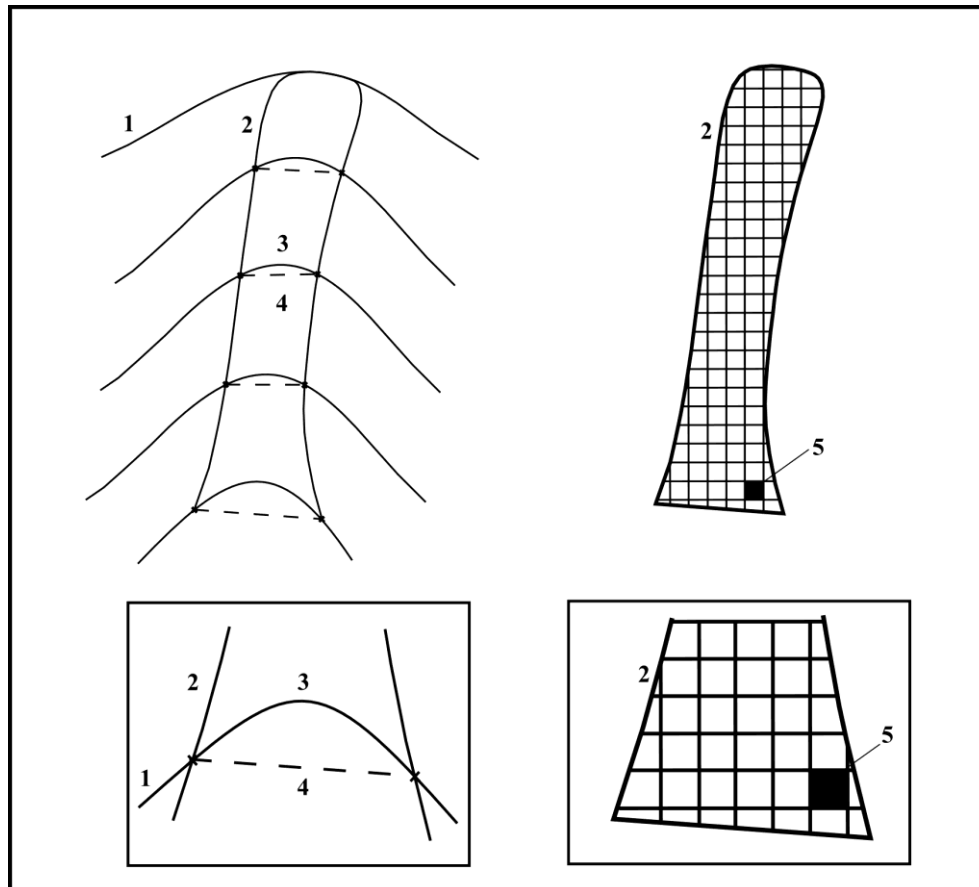


Fig. 3. Components of the contour line section index (CLSI) and charachteristic inclination of valleys methods.
 1 = contour line; 2 = border of the derasional valley according to the geomorphological map; 3 = contour line section between the incoming and outgoing point of the contour line; 4 = straight line between the incoming and outgoing point of the contour line; 5 = one pixel of the derasional valley, which represents the slope category of the derasional valley.

The data related to each derasional valley were organized into a chart: the elevation values of the contour lines crossing the valley were combined with the length of the given contour line in the valley, as well as with the length of the straight line between the incoming and outgoing point of the contour line. From the quotient of these two values, the contour line section index (CLSI) belonging to the contour lines crossing the valleys was calculated. This process was repeated in the case of every valley. The CLSI characteristic for each valley was added and averaged, and their standard

deviation was defined. The CLSIs were illustrated on graphs, so they became comparable with each other.

In the second stage of the research (second method), the characteristic inclination of the valleys was analyzed. Using the *r.slope.aspect* routine, the slope map of the Somló Hill was constructed. This routine used to determine slopes uses a 3x3 neighbourhood around each cell in the elevation file (Shapiro et al., 2006). The slope map was reclassified using the *r.reclass.rules* routine, so that the slope angles were grouped every 3 degree.

(Therefore the area with a gradient between 0 and 3 degrees got an identification number 1, those with a gradient between 4 and 6 got an identification number 2 etc.).

From the geomorphologic map, the derasional valleys were distinguished using the *r.reclass.rules* routine, so that they should get the value 1 during the reclassifying process, while the other landforms got NULL (= no data). With the use of the *r.clump* routine, different values were connected to each valley. From the resulting map the valleys were distinctly selected using the *r.reclass.rules* routine. These were multiplied one by one with the already described slope map, using the *r.mapcalculator* routine. The maps representing the slope categories of each valley were reported back using the *r.report* routine, so that the area (in hectares) connected to the given slope category as well as the territory of the valley were saved into a file.

For every valley, the ratio (in percentage) of each slope angle related to the area of the valley, their average value, as well as their standard deviation were defined. The results were compared with a normal distribution, and they both were displayed on a single graph.

Results

In the course of the research, the CLSIs of 19 young and 6 old valleys were examined (Fig. 4). The investigation of the young valleys numbered 17, 18 and 22 could not be completed due to their special morphologic location. The defile running in valley Id. 22, as well as the incorrect location of the valleys Id. 17 and 18 on the geomorphologic map could mislead the results of the measurements and also the consequences related to the young derasional valleys.

<i>CLSI of young derasion valleys</i>				
Valley ID	Min.	Max.	Average	Standard dev.
1	0.43	1.02	0.77	0.16
2	0.68	1.02	0.89	0.1
3	0.74	0.98	0.88	0.07
4	0.67	0.98	0.78	0.07
5	0.81	0.98	0.92	0.05
6	0.88	0.99	0.95	0.03
7	0.88	1	0.95	0.04
8	0.83	0.96	0.91	0.04
9	0.88	0.99	0.94	0.03
10	0.79	1	0.93	0.07
11	0.54	0.99	0.89	0.12
12	0.90	0.98	0.95	0.03
13	0.79	1.02	0.93	0.06
14	0.68	0.98	0.89	0.07
15	0.64	1.00	0.83	0.84
16	0.58	1.01	0.83	0.12
19	0.87	1.04	0.94	0.05
20	0.81	1.03	0.93	0.05
21	0.67	1.00	0.81	0.08
Average	0.74	1.00	0.89	0.11
Standard dev.	0.12	0.02	0.06	0.18

<i>CLSI of old derasion valleys</i>				
Valley ID	Min.	Max.	Average	Standard dev.
1	0.30	1.01	0.71	0.23
3	0.44	0.98	0.74	0.17
4	0.61	0.96	0.87	0.09
5	0.95	0.99	0.98	0.01
6	0.39	0.93	0.67	0.16
Average	0.54	0.97	0.79	0.13
Standard dev.	0.26	0.03	0.13	0.08

Fig. 4. Common CLSI of the young and old derasional valleys

The CLSIs of the contour lines present in the young derasional valleys vary between 0.43-1.04, while that of the old derasional valleys have a value between 0.07-1.01. Based on the average value of the contour lines in the valleys, the distinct CLSIs characteristic for each valley have a value of 0.89 in the case of young derasional valleys, accompanied by a low standard deviation of 0.11. This value becomes 0.79 in the case of old derasional valleys, with a standard deviation of 0.17. In the case of young valleys, the above mentioned value primarily resulted from the CLSIs above 0.8; lower values occurred only in the valleys Id. 1 and 4. The old valleys have a CLSI below 0.8; the valleys Id. 4 and 5 represent exceptions.

Because of the spatial restrictions of the study, there is no possibility to introduce the CLSIs of every valley; therefore the results of the investigation are exemplified on some characteristic instances. The morphologic features of the young valleys are best represented by the derasional valley Id. 20 (Fig. 5): its high average CLSI (0.93) is accompanied by a standard deviation of 0.05. The CLSIs of each contour lines vary between 0.81 and 1.03.

Among the CLSIs characteristic for the young valleys, the indexes of the valley Id. 1 (Fig. 6) did not fit into the line: their average value is 0.77 and the standard deviation 0.16. Regarding the CLSI-graph of the valley, it becomes evident that below and above the altitude of 200 m the indexes show a lower standard deviation, according to which two parts of the valley could be differentiated: the section below 200 m was named 1A, while the section above 200 m was named 1B. The average CLSI of the new valley section became 0.86 in the case of 1B, which is characteristic for the young valleys. The validity of the differentiation is also proven by the outcome of a low standard deviation. The 1A valley section has features of old valleys, though its values are even lower: the average CLSI is 0.61; while the standard deviation is 0.14.

The average CLSI (0.87) and standard deviation (0.09) of the old derasional valley Id. 4 is not in compliance with the typical values of old valleys (0.79 average of CLSI, standard deviation above 0.1). Nevertheless, with the graphic representation of every index of the valley (Fig. 7), a tendency of the CLSIs became obvious, which unambiguously groups them into this category. The values of the CLSIs continuously decrease from 380 m to 360 m, then take the value of 0.6, and increase again towards lower altitudes. The values of the valley above 380 m equal that of the young derasional valleys. The average value (0.98) of valley Id. 5 fits the least into the CLSIs of the old derasional valleys (Fig. 8). This value is accompanied by an exceptionally low standard deviation of 0.01.

The investigation aiming at the exploration of the slope categories of the valleys was completed in all the cases of the 22 young and the 6 old valleys. The graphic representation of the normal distribution of the young derasional valleys (Fig. 9) made it evident that nearly all of them show a distorted value. The valleys were characterised by maximum values of 0.0418-0.0254, and by a low standard deviation (0.01). Only the valleys Id. 5, 10 and 17 meant an exception.

The valleys Id. 5 and 10 were exceptional because of their low maximum value (0.0121) and the distinctively low standard deviation (0.003 and 0.004). The young valley Id. 17 showed an atypical picture as well, with a maximum value of 0.0535; and a standard deviation of 0.02. Real normal distribution was detectable only in this valley.

In the case of old derasional valleys, every valley showed a distorted normal distribution (Fig. 10). The maximum values taken by the curve varied between 0.0536 and 0.0418, their standard deviation was 0.02 – except for the valleys Id. 2 (maximum value of 0.0321; standard deviation of 0.01) and 5 (maximum value of 0.0804; standard deviation of 0.01).

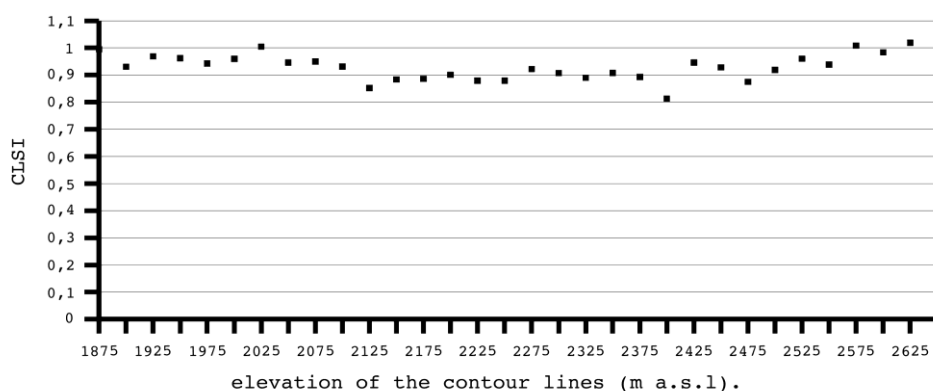


Fig. 5. CLSI of the young derasional valley Id. 20

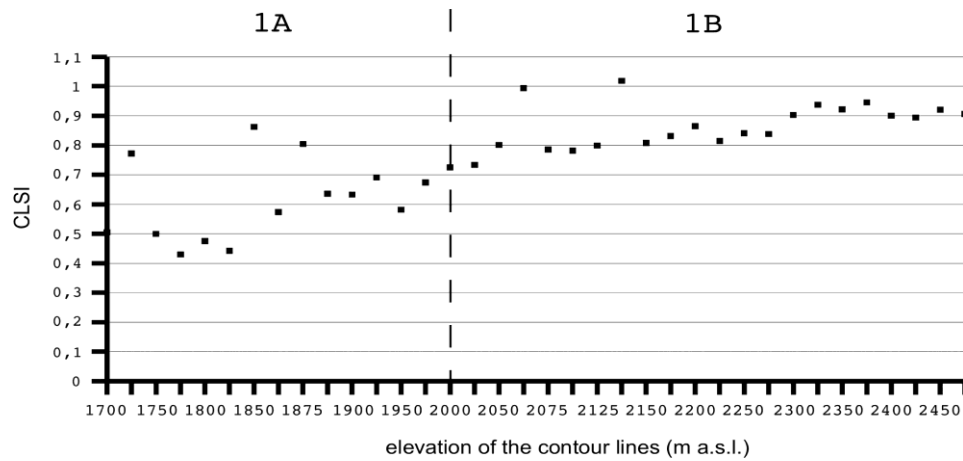


Fig. 6. CLSI of the young derasional valley Id. 1

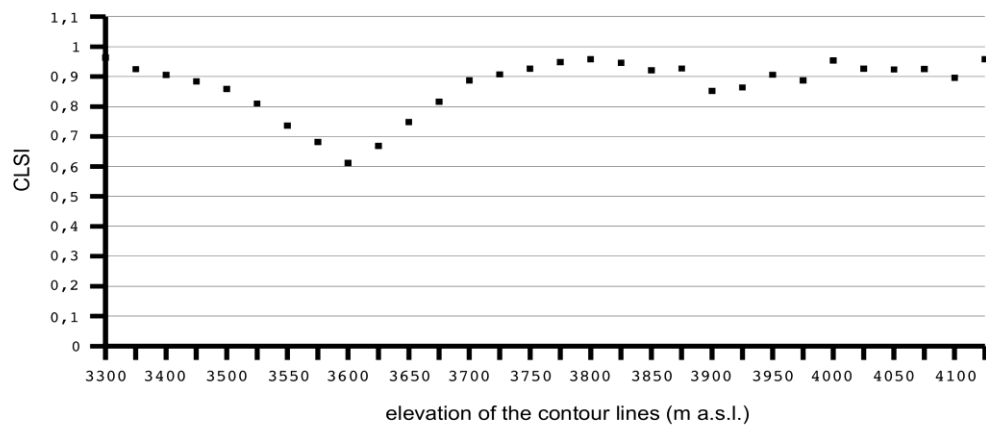


Fig. 7. CLSI of the old derasional valley Id. 4

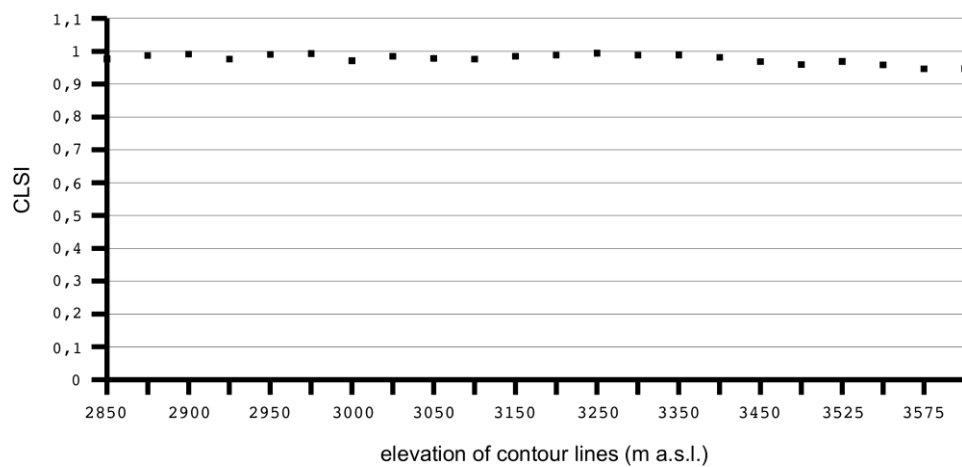


Fig. 8. CLSI of the old derasional valley Id. 5

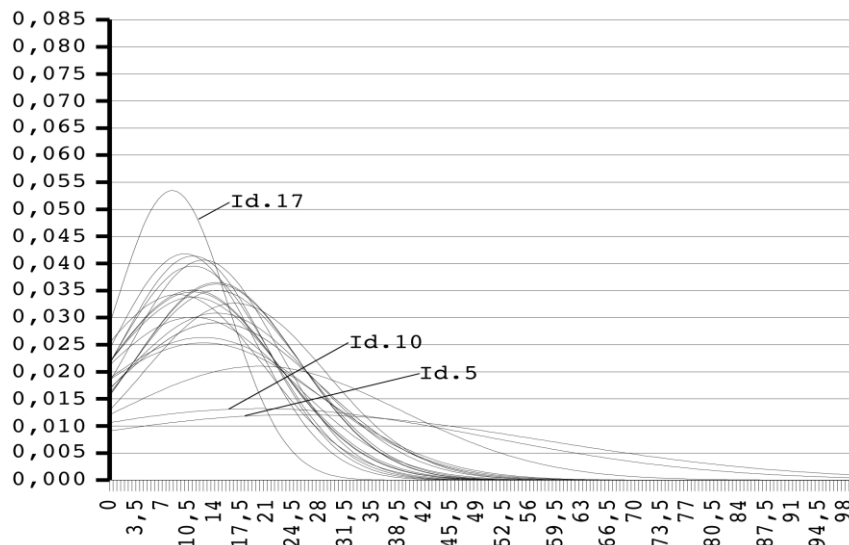


Fig. 9. Normal distribution of the slopes of the young derasional valleys

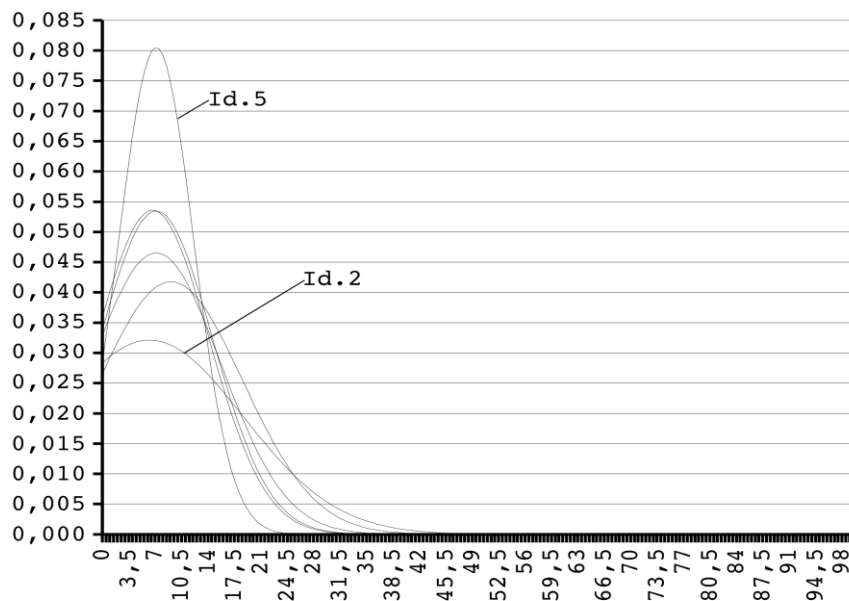


Fig. 10. Normal distribution of the slopes of the old derasional valleys

Discussion

The CLSIs are the index numbers of the cross-section of the valley. The longer the contour line section related to its chord, the lower is the CLSI. The lower the CLSI at a given point, the larger is the determined cross-section of the valley. According to the factors of the index, the shape of the valley (V-shaped or bowl-shaped) can not be defined in the research area, only its extension can be concluded.

The CLSIs observed in the valleys and having a value 1 or more mean, that at the given valley section, the length of the contour line in the valley

equals or is shorter than the length of the straight line between the incoming and outgoing point of the contour line. The latter unambiguously refers to a measurement mistake. In the former case, we can realize a completely plain territory. Such sections can be observed only at the lower or upper ending of the valleys. The mistakes occurring at the construction of the geomorphologic map – primarily during the generalization process – as well as an inaccurate manual measurement can result in such an error. Therefore, either these sections are not part of the valley – thus the geomorphologic map is incorrect –, or the sections are part of the valley, but

the measurement is mistaken. The above mentioned values rarely occur (only once in the case of old derasional valleys), so they did not influence the precision of the statistical connections. The value of 0.43 is considerably low, which refers to a valley section of a significant extension. Similarly low values occur mainly in cases of old valleys.

The young valleys with a high CLSI (above 0.8) and low standard deviation (0.11) can be well differentiated from the old derasional valleys, which are characterized by a low CLSI (below 0.8) and a high standard deviation. The young valleys are plain, bowl-shaped forms, typical derasional valleys, which are distinguished by low CLSIs. The high CLSI and its low standard deviation value proves the fact that the valleys are even, longitudinal hollows, in which – besides some exceptions – sections with significantly different values do not, or just rarely occur. Among the young valleys, Id. 20 best exemplifies this type of valley. The already mentioned differences were revealed by the examinations; therefore their accurate location is also described.

Anomalies of this kind are observable in the case of the young valley Id. 1. Based on the CLSIs the valley was divided into two sections. The distinct valley sections show more consistency individually – this fact is also proven by the low standard deviation of section 1B. The discussed section took the value characteristic for young valleys. Section 1A has CLSI values even lower than the old valleys, and its standard deviation is high as well. According to the results of geomorphologic mapping (Kovács, 2008) it can not belong to the old valleys, thus its formation can surely be dated back to the Pleistocene (i.e. it is a young valley). Nevertheless, the shape of the valley can be significantly influenced by the rock quality as well as by many other factors. Therefore the described method is able to reveal the extent and location of the anomalies to be found in the cross-sections of valleys, though it can not inform us about their origin and accurate characteristics.

The old valleys are distinguished by significantly lower average CLSIs (below 0.8) and higher standard deviations (0.13). The high CLSIs can be explained by the relatively deeper valleys. The high standard deviation of the valleys refers to their varied morphology. This diversity, which is observable in almost every case of old valleys, is especially noteworthy on the graph of the old valley Id. 4. The CLSIs reach their minimum after continuous decrease at the lower section of the valley, and then start to increase again, until they take the value characteristic for young valleys. This

phenomenon is also explained by the results of the traditional geomorphologic mapping. The old derasional valleys occur at their upper sections as plain, bowl-shaped hollows. Their CLSIs almost exclusively equal the values of the young valleys. At lower altitudes, reaching the basalt mesa, the valleys turn into V-shaped forms, and even incise into the basalt nappe. The decrease of the CLSI noteworthy in the case of the old valley Id. 4 is caused by this valley section and by the incision itself. Thus the method revealed a considerable morphometric change and also its location in the case of old valleys. Nevertheless, one fact cannot be ignored, namely that the low CLSIs measured in old valleys (plain, bowl-shaped upper valley section) equal the values of the young valleys, though have fundamentally different genetics. So the research found equal results for two formations of essentially diverse genetics, but of analogous morphologic parameters.

Similar results occurred in the case of the old derasional valley Id. 5 as well. The average CLSI of the valley is 0.98, which refers to an even more insignificant form than the young valleys have (very mild, shallow surface hollow). Its standard deviation is considerably low (0.01), therefore the valley is consistent and there are no detectable morphometric changes. According to the field researches – and despite its present-day morphologic parameters – it is definitely an old valley, filled up and nearly completely covered by the debris originating from the basalt nappe, which caused its shallow form.

The investigation of the normal distribution of the valleys' slopes was carried out in each valley. Based on the maximum values of the normal distributions, the categories of young and old valleys could be constructed once again. The maximum values of the young valleys varied between 0.0418 and 0.0254, while those of the old valleys between 0.0418 and 0.0536. Related to the defined value limits, some valleys showed differences again.

The valleys Id. 5 and 10 were distinguished based on their lower maximum and extremely low standard deviation. Their values moved on a wide range, several values occurred with a low frequency. The valley Id. 17 is prominent among the other valleys because of its outstanding maximum value and real normal distribution. The comparison with CLSI researches is possible only in the cases of valley Id. 5 and 10, for valley Id. 17 it could not be examined. The valleys at that location did not show any difference from the average values.

Compared with the maximum values of the old valleys, Id. 2 and 5 had lower values. The

exceptional valley Id. 5 was already realized during the former CLSI-investigation, since it had values characteristic for the young valleys. Its high maximum value can be explained by its filled up position (significantly even, shallow bowl-shape).

The research was also able to distinguish two types of derasional valleys. Nevertheless, it only provides little basic information about the morphometry, and tells nothing about the genetics. The comparison of the results of the two investigations showed only one coincidence.

Summary

The average CLSIs, as well as the standard deviation of both young and old valleys were defined, according to which it was possible to distinguish the members of these two groups, and to list the morphometrically different valleys. As the most characteristic example for young valleys, valley Id. 20 was introduced. In the case of the old valley Id. 4, the method revealed the exact location of the dissimilar valley sections and the extent of divergence.

The average CLSI of some valleys differed from the average value of the two groups. The separated young valley Id. 1 could be divided into two valley sections according to its CLSIs. In the case of the old valley Id. 5, it was achievable to reveal values different from the average.

The CLSI served as a kind of index characteristic for the cross-section of the valley, based on which it was possible to specify the changes and differences from the average in the cross-sections of the valleys. The index provided

information about their proper location, and also about the extent of change. Nevertheless, it is incapable to define the shape of the given valley section, or to reveal the reasons for the development of different sections. Therefore, it is necessary to continue the field researches, as well as to apply the additional means of the traditional geomorphologic mapping.

The analysis of the slope categories of the valleys represented a novel point of view in this study. According to the maximum values of the normal distribution projected upon the slopes of the valleys, it was also made possible to distinguish the young and old derasional vales. The study exposed the valleys with values different from the characteristic features of these two groups (young valleys Id. 5, 10 and 17; old valleys Id. 2 and 5).

The comparison of the results of the two methods showed agreement in the case of the old derasional valley Id. 5, thus the necessity of its further investigation was pointed out by both methods simultaneously. Therefore, these techniques are able to complete the comparative and analytic investigation of basic morphometric features of valleys, as well as to assist the geomorphologic mapping.

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Morphostructure and Morphodynamical Processes in the Milcov Morphohydrographic Basin

Florina GRECU, Răzvan SĂCRIERU

Key words: vulnerability, Milcov, landslides, longitudinal faults, transversal faults, lithological contacts

Abstract. The present work aims at analyzing mainly the vulnerability to the imminent tectonically induced geomorphologic risk. The Milcov morphohydrographic basin is situated at the exterior of the Bend Zone – a highly tectonically active region. Most part of the observed area is of medium and high vulnerability in the present surface relief modelling processes, and tectonics is in a greater degree a preparatory factor than a triggering factor. In the internal Subcarpathian sector, major reactivation of landslides occurred along longitudinal and transversal faults. The slopes are in continuous imbalance. The geological structure mirrors well the morphology of the landslides. In the medium sector, landslides occur mainly on lithological contacts. The inferior sector shows increased vulnerability to hydric and river bed processes.

Rezumat. Lucrarea de față își propune să analizeze cu precădere vulnerabilitatea la riscul geomorfologic iminent indus tectonic. Bazinul morfohidrografic Milcov este situat în partea externă a Curburii Carpatice – o regiune deosebit de activă din punct de vedere tectonic. Cea mai mare parte a arealului analizat se încadrează în clasa de vulnerabilitate medie și mare la procese de modelare actuală iar tectonica reprezintă în primul rând un factor pregătitor și mai puțin unul declansator. În sectorul subcarpatic intern importante reactivări ale alunecărilor de teren s-au produs în lungul faliiilor longitudinale și transversale. Versanții se afla permanent într-o stare de dezechilibru. Structura geologică se reflectă foarte bine în morfologia alunecărilor de teren. În sectorul mediu alunecările de teren apar în special pe contacte litologice. Pentru sectorul inferior s-a observat o vulnerabilitate mare la procese hidrice și de albie.

Introduction

The Milcov morpho-hydrographic basin develops outer area of the Bend Zone. Within the Milcov river is a tributary of the Putna river (Fig. 1). The longitudinal profile extends over 84 km, between 1040 m altitude (springs), and 25m altitude to the Rastoaca confluence (Eastern from Focsani). The geographical situation in the most active tectonic area in Romania and the most seismic active area in Continental Europe, as well as other external factors, is the condition for an intense geomorphological processes pattern.

The present work aims at analyzing mainly the vulnerability at the geomorphological, tectonically induced imminent risk. We thus evaluate the existence of vulnerability and, respectively, natural risk (Grecu, 2006, 2007) or biophysical vulnerability, which is the response of certain systems to the alterations of other systems from the environment, which they come in contact with

(Birkmann, 2006, World Conference on Disaster Reduction; Hyogo Framework for Action 2005 – 2015”) (Grecu, 2009).

The tectonics and the stratigraphy of the basin (alternating pervious and impervious rocks) decisively influences the dynamics of the relief. Geologically, the Milcov basin and the entire Bend Zone have been subject to several research trends: the geological general analysis (Mateescu, 1927) the geology and the tectonics of the Neogene molasse (Pauca, 1942, 1952), the Milcov Strata (Macarovici, Motnas, Contescu, 1967), the geology of the Neogene deposits between the Milcov Valley and Putna (Cehlarof, 1998), the geological development starting with the Pliocene (Necea, Fielitz and Matenco, 2005), the tectonic characteristics of the Vallachian Phase (Hippolyte, Săndulescu, 1996), the subsidence rate of the Focsani Depression (Tărăpoancă, 2003), the present subsidence from GPS measurements (Hoeven van der et al., 2005).

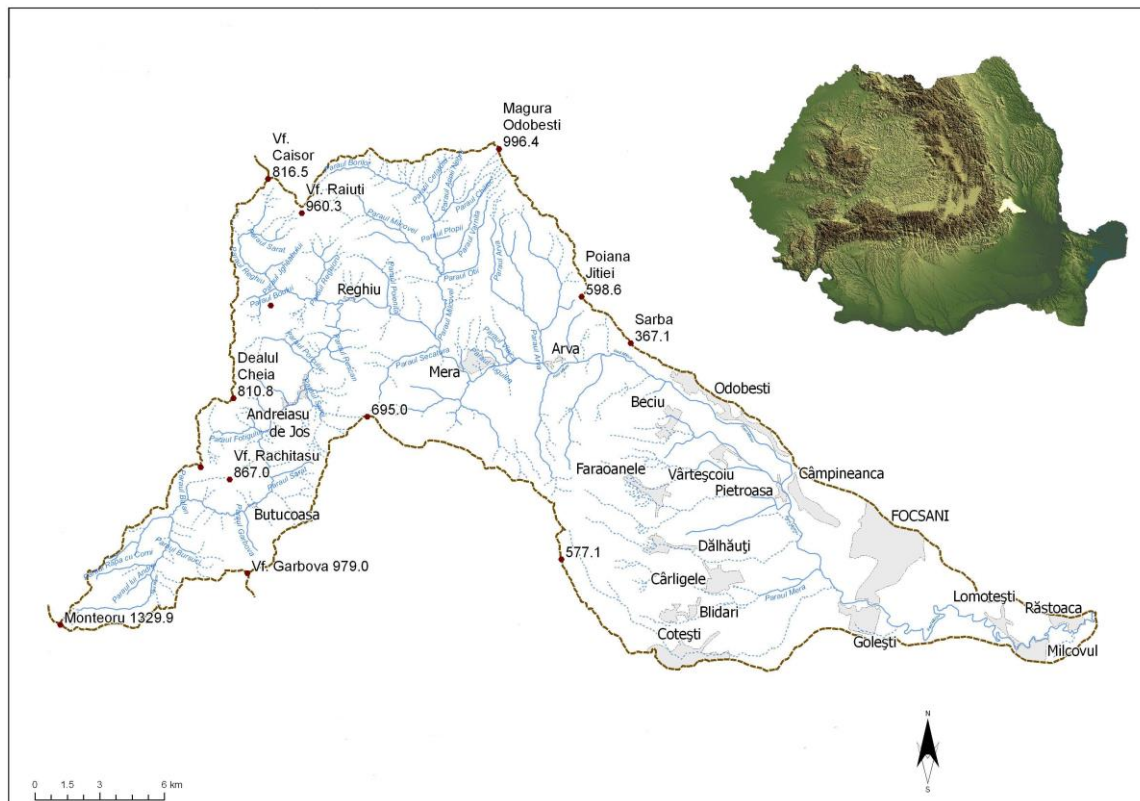


Fig. 1. The Milcov hydrographic basin. Geographical setting in Romania

The basic method is the geo - morphological one, field observation and mapping using topographic, geological or tectonic maps. Furthermore, in order to determine the degrees of vulnerability, special analytical, geomorphic (morphometric and morphographic) maps have been drawn.

The geological structure and the tectonic evolution as factors of instability

The present day morphodynamics is a consequence of the tectonical mobility of the region, manifested through plicative movements and earthquakes. Most of the analyzed area is placed within the high or medium risk class vulnerability to present shaping processes.

Structurally, it comprises three units: Paleogen flysch, Neogene molasse, and the foredeep area. The Paleogene flysch (marls, limestone grits, Kliwa sandstone, menilite schists and dysodile schists) appears as four anticlines on Eocene formations and two Oligocene synclinal bands alternate with the anticlinal Eocene bands (Mateescu, 1927). The Neogene molasse (sandstone, salt, saltbreccias, marls) are featured by the formation of thrust faults and the salt dyapir (Reghiu, Andreiasu). There are

also accumulations of natural gas in correspondence with the Casin-Bisoca Fault (The alive fires at Andreiasu). Five folds within the Pre-Sarmatian Miocene formations have been identified (Paucă, 1952). The foredeep area is situated Eastern of the Cașin – Bisoca fault and comprises the Milcov Strata (Sarmatian – Pliocene) which have a flysch configuration due to the sedimentation speed being the same as the subsidence speed (Macarovici et al., 1967).

Tectonics are primarily a preparatory factor, rather than an initiating one. Earthquakes of over 6 or 7 degrees are the exception (ex. 1940, 1977, 1986, 1990) which prodecod the landslides in Andreiasu de Sus and Titila area. The most significant reactivations have taken place after rainy periods (for example, July 2005). The tectonic factor played an essential part in relief evolution through the shaping of the hydrographic network and the marked slope dynamics.

The Neogene molasse englobes a series of crushed synclinal (Raiut-Titila, Ursoaia-Rachitasul) with a tendency towards scaling (Paucă, 1952). Locally there are dyapir salt structures (Andreiasu, Reghiu). Such a complex tectonic structure is a consequence of longitudinal and transversal faults, appeared since Moldavian (Miocene) and Valachian

(Pleistocene) tectogenesis. Major faults (e.g. Cașin – Bisoca) have formed during the Miocene. The high number of faults is due to the modification in the stress field. The direction of compression changed from NW-SE to N-S during middle-Tortonian time (Hippolyte & Săndulescu, 1996).

The longitudinal faults in Jghiaburi – Butucoasa, Lunca, Prahuda area are reversed faults oriented from North to South. The transversal faults are mainly strike-slip faults and are oriented from West to East and are less evident: the Raiuti Fault, the Farcasei Fault, the Andreiasu Fault, the Reghiului Fault, the Titila Fault. These lead to a discontinuity in the linear-longitudinal relief structures and the formation of smaller and dynamically different compartments.

The Neogene molasse comprises Miocene formations between the marginal fault of the flysch in the west and the Cașin – Bisoca Fault to the east. This includes the Aquitanian (salt and salt breccias), the Burdigalian (mainly sandstone and marls), the Badenian (the Rachitasu sandstone, grey marls, the Slanic ash tuff) having a lithology with a high potential for slope processes.

The vulnerability to geomorphological processes

The geomorphological processes with direct and rapid response to earthquakes are the gravitational ones.

Concerning slope instability we observe:

- the occurrence of microlandforms and old material deluvial, which are disturbed and reactivated even by minor earthquakes;
- the landslides movement or the separation of material masses from the slope following seismic shock.

Both situations are significantly influenced by the restrictive factor of the vegetal cover features.

The most affected area by slope processes within the neo-tectonic and masive deforestation conditions is the upper Milcov basin, upstream from the Milcov-Reghiu confluence. The distribution of landslides depends on the Milcov Valley, as well as the tributary basins.

The Milcov Valley in this internal Subcarpathian sector follows two major longitudinal fault lines: Jgheaburi – Butucoasa and Casin – Bisoca. The slopes of the Milcov Valley are affected by old and deep landslides (Fig. 2). Reactivations of such landslides may be observed on the two slopes of the Milcov River between Rachitasu and Andreiasu de Jos. Further there also appear gully processes (left slope). Subsequent to the year 1960 the versants have been protected by forests (black

pine, silvester pine) which presently have reached their maturity. At the same time, transversal hydrotechnical works has been conducted on the majority of affluent courses, thus ensuring the slope stability (Săcrieru, 2008 a).



Fig. 2 Landslides on the Cașin Bisoca fault
(Photo: Grecu, September 2008)

Most part of the lower basin of the Roșoiu stream is affected by the landslides along two longitudinal faults. The land degradation is also related to gullies. Due to ample forest and hydrotechnical works subsequent to the year 1963, we estimate a diminished vulnerability to landslide processes.

In the Upper basin of the Pârâu Sării, within the areas of Andreiașu de Sus and Titila (Fig. 4), we estimate a high vulnerability to landslides due to the placement between two important tectonic lines: the Cașin Bisoca fault (east) (Fig. 3) and the axis of the Răiuț Titila synclinal (West). The upper course of the Pârâu Sării developed along the reverse fault of Andreiașu and landslide occurred on the entire slope. Thus the areas between built area and also the arable land, orchards, pastures and hay-fields. On the lower course the marked instability is due to the presence of a salt dyapir. This featuring the areas where reactivations of landslides occurred after July 2005. Also there have been mudslides upstream from the confluence with Milcov, at Eastern of Andreiașu de Jos.



Fig. 3. The Milcov Valley along the Casin-Bisoca fault
(Reghiu-Scruntar) (Photo: Grecu, September 2008)



The Lingura Valley (Păpăluga) catchment is mostly affected by landslides. These are complex and deep landslides. Between the landslide ridges there are microdepressions with lakes and small surfaces of water, mostly silted. A landslide (Fig. 5) occurred in the springs area along the Andreiașu transversal fault and the area of separation near the Cașin Bisoca fault.

We assess the Andreiașu de Jos – Focul Viu area as a particularly mobile area, especially due to the intersecting of the longitudinal faults with the transversal Andreiașu fault. Thus slopes and

riverbeds are dynamic state (Fig. 6 and 7). This mobility is accentuated and permanently maintained by frequent earthquakes (sometimes daily). As regards of small hydrographic basins the landslides on both slopes obturate riverbeds transforming them into successions of pools and incised thalwegs. There are numerous examples, both in the upper and middle basin. The disafforested area on the left side of the Milcov is the most affected.

There appears consequently the permanent action of earthquakes on the slope morphodynamics with paroxysmal pulsations at high magnitude earthquakes.

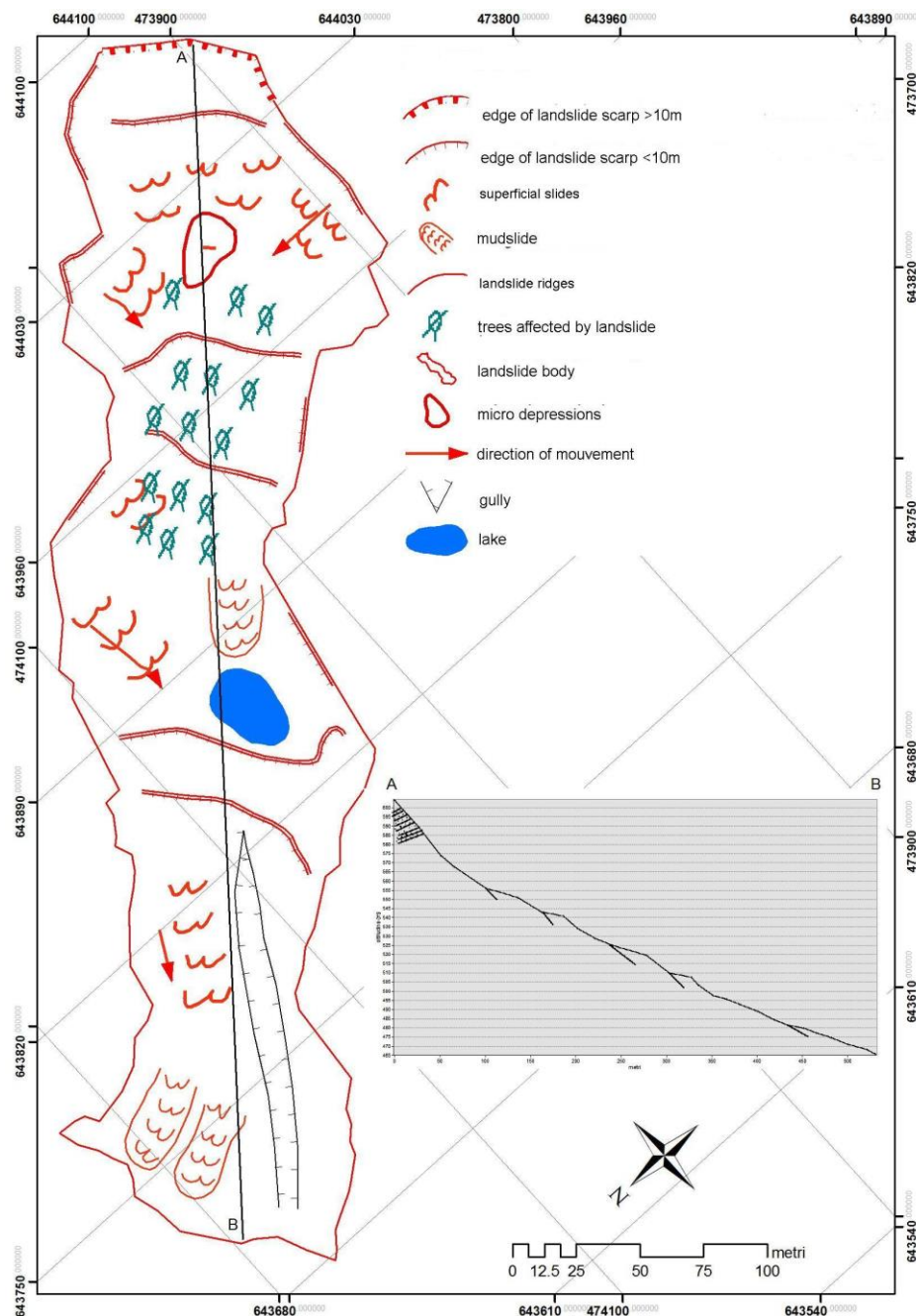


Fig. 5. The Păpăluga slide. Geomorphologic outline and longitudinal profile (Săcîriu, 2008 c)

The medium sector (the Mera depressions and the external subcarpathian hills), is featured by numerous lithological contacts and a density of over 3 km/km² in the upper catchments and extended areas along the Milcov, show a medium and high vulnerability to slope processes. In numerous situations landslides have advanced regressively on the slope, affecting the interfluvies, the watershed being under the initial one. This is the case of the afluent of Milcov on the right, especially the southern watershed of the Tulburea Valley, afforested slopes.

For the Mera depression the lithological factor is essential in initiating landslides. The Upper Pliocene formations are the most prone to mass slides on versants. The overhumid clays, highly contractile, with a low resistance to shearing, become a lubricant determining landslides on strata with less than 10° inclination. Due to the high hydrostatic pressure of underground waters, they engage fine particles of sand, determining a suffusion process followed by complex landslides. (Zamfirescu et al., 1975).

The lower sector, of glaciis type and plain, is mainly affected by hydric and riverbed processes. Upstream from Câmpineanca the Milcov river meanders significantly, as a consequence of active subsidence (Grecu et al, 2007).

Floodplain processes activated after the increase in discharge from 12-18 July 2005 manifested differently from one sector to another. Around Odobesti town, the lateral erosion is the geomorphological with the greatest impact at the riverbed level (Săcrieru, 2008 b). In the lower sector, the alluviation from the minor riverbed has exceeded tens of centimeters. In Golești, for

example, a 1,2 m stratum of fine alluvions was deposited, at an absolute maximum debit of 724 m³/s. Flooding has produced significant material damage, especially along the access ways and national roads.

Conclusions

The natural hazards within the Milcov basin are usually complex, the triggering factors being of geological nature (neotectonical, seismic, petrographic) and climatic, to which we can add deforestation and other anthropic factors.

Earthquakes are a permanent cause of the geomorphologic processes because they gradually decrease the slope stability.

Depending on the intensity and frequency of the processes affecting the risk elements, the vulnerability is medium to high on the Milcov catchment, with local differences. Generally it is a adjacent risk, with an occurrence probability dependent on earthquakes of more the 6 degrees Richter, but with continuous quantitative accumulations.

The correlation between the prediction and the occurrence of earthquakes is extremely important on the reduction of the impact on the population, through the possibility of informing. The estimated magnitude and the present day earthquakes in Vrancea in February 2005 show us a good prediction, usually the actual magnitude being lower than the predicted one. Comparatively, for the earthquakes from the seismic areas of Europe and Asia, the correlation degree between the figures mentioned in February 2005 is over 0,70 (Grecu, 2006, 2008).



Fig. 6. Slided rock son a fault slope
(Photo: Grecu, September 2008)



Fig. 7. The influence of microstructure in slope drainage system. Reghiu (Photo: Grecu, September 2008)

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Present-Day Geomorphological Processes and Geomorphological Risk in the Sibiu-Apold Depressionary Passageway (Sw Transylvanian Tableland)*

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Key-words: geomorphological processes, geomorphological risk, Transylvanian Tableland.

Abstract. The contact area between the Southern Carpathians and the Transylvanian Plateau is marked by a number of depressions, between the valleys of the Olt in the east and the Sebeş in the west: Sibiu, Sălişte and Apold, a contact depressionary passageway (80 km long and 700 sq.km). The Sibiu-Apold depressionary passageway is characterized by a great morphological complexity. This complexity results, on the one hand, from the position and the location of a chain of depressions at the morphostructural contact between the two units and, on the other hand, from the division into two hydrographic basins: of the Mureş River, through the Secaşul Mare, and of the Olt River through the Cibin, which have distinct drainage conditions and underwent major hydrographic changes in the Quaternary, very likely in its early stage.

The Sibiu – Apold depressionary passageway is affected by a great diversity of geomorphological processes, the frequency and magnitude of which could be better explained if viewed within the framework of global environmental change. Quantitative evaluations of the present-day mass movements, and gully erosion show significant regional differences which relate to recent extreme climatic events and land use changes.

Therefore, the very dynamic character of slopes is a major characteristic of the Apold Depression, which represents the western sector of the passageway. It covers one third of the passageway (the sector connected with the lower base level of the Mureş River). On the other hand, in the Sălişte and the Sibiu depressions, which represent the central and the eastern sectors of the passageway (connected with the Olt Basin), the relief is insignificantly affected by denudation, most of the slopes being in a very stable state. In general, instability on slopes produces landslides associated with gully processes. The extension, frequency and intensity of present-day geomorphological processes depend on such factors as precipitation, categories of slope and human activity (deforestation, land uses orchards, crops).

The frequency and intensity of present-day geomorphological processes have led to the identification of three classes of geomorphological risk: high, moderate and low.

Present-day modelling and the factors influencing it

Present-day geomorphic processes are triggered either by general factors (affecting all of the Sibiu-Apold depressionary passageway), or by local ones. The contact area between the Southern Carpathians and the Transylvanian Plateau is marked by a number of depressions, between the valleys of the Olt in the east and the Sebeş in the west: Sibiu, Sălişte and Apold, a contact depressionary passageway (80 km long and 700 sq.km) (Fig. 1). The Sibiu-Apold depressionary passageway is characterized by a great morphological complexity. This study is the result of field investigation, as well

as of the correlations established among several factors: climatic (with focus on the precipitation regime - 600-700 mm/year), geological, morpho-hydrological (surface and underground waters, categories of slope, aspect and length of slopes, relief energy, density of relief fragmentation), biological (degree of vegetation coverage) and anthropic.

Climatic factors

The climate of the depressions is moderate continental with temperate shades, the geographical position of these depressions accounting for a greater opening towards the north-west and west

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increases the circulation of the masses of maritime air from the west. So, in summer, the weather is unstable, cool, with much nebulosity and rainfall, in winter temperatures and air moisture register higher values (Bogdan et al., 1981).

Local factors affect differently the main variables registered by the three depressions in terms of the characteristic relief of each of them. Mean air temperatures in the December-February period are negative (with -4.2°C in Sibiu and -3.7°C

C in Sebeş in the coldest month); from March through to November the mean temperatures are 3.5°C and 23°C respectively, top values being registered in July (19.4°C in Sibiu and 19.9°C in Sebeş). The lowest annual temperatures may drop below -30°C with an absolute record of -31°C in Sibiu and -30°C in Sebeş. The highest annual temperature may exceed 30°C and so may the absolute value: 37.4°C in Sibiu on July 3, 1952.

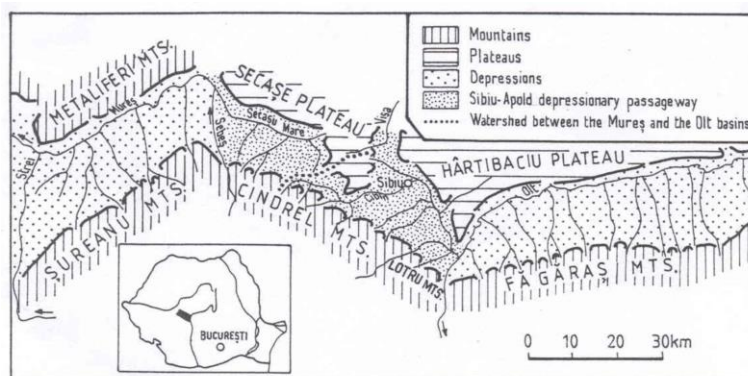


Fig. 1. The Sibiu-Apold depressionary passageway between the Olt Valley in the east and the Sebeş Valley in the west

Freeze-thaw phenomena play a secondary role in present-day modelling processes, occasionally influencing creep and piping on the lefthandside slopes of the Mag and the Cibin valleys (between the settlements of Guşteriţa and Tâlmăciu on the Cibin) and on the righthandside slope of the Secaşul Mare (between Apoldu de Jos and Sebeş), etc.

It is precipitation that enhances and accelerates those processes. Mean quantities go up to 600-700 mm/year, and even higher: Sibiu 653.4 mm/year, Sălişte 697.4 mm/year and southwards, at the contact with the mountain (817 mm/year at Boiţa, 781.2 mm/year at Răşinari, etc.), but decreasing northwards, on the contact line with the tableland (Şura Mică 613 mm/year, Apoldu de Jos 571.6 mm/year and Sebeş 568.7 mm/year. The wide monthly, seasonal, annual and multiannual

variations, visibly influence the relief modelling regime. Maximum rainfall/24 hrs occurs in July and August (between 40-80 mm), highest quantities being registered in the warm season (April-September) – with a maximum in June (over 100-120 mm), and occasionally in July. Unperiodical climate variations registered excessively rainy years (1966, 1969, 1970, 1975, 1991, 1997, 1998, 1999, 2005), followed by the reactivation of landslide and erosion, very significant in some cases. A very important element in unleashing them is the substrate.

A synthesis of the main climatic variables is based also on the Péguy climograms (Fig. 2) revealing the effect on modelling processes of snow which, accumulating in the cold and humid season begins to thaw and flow as temperatures suddenly rise during the warm humid months.

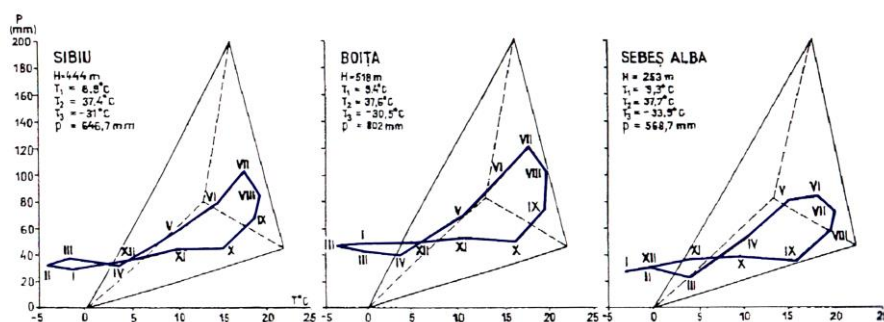


Fig. 2. Péguy-type climograms: H, absolute altitude. Temperature: T1, annual mean in the air; T2, abs. max; T3, abs. min; P, annual amount of precipitation

Morpho-hydrographical factors

One of the main factors influencing the intensity and rate of relief modelling (slopes, channel-beds) is the discharge regime. Notable differences in this respect occur between torrential basins and permanent basins. Thus, the Secaşul Mare cuesta, the scarps on the lefthandside of the Mag and the Cibin valleys (between Guşteriţa and Tălmăciu) are cut by small 2nd and 3rd-order basins that have a pluvial regime. The 5th and 6th – order basins (Horton-Strahler classification) – the Sălişte, Sibiel, Seviş, Cislădie, Sadu, etc., which cross the three depressions, have a pluvial-nival regime. The increased discharge and suspended sediment load on these watercourses across the three depressions is different: an average multiannual value of 4.72 cum/sec for the Cibin and only 2.37 cum/sec for the Secaşul Mare.

Looking at discharge rates by season, differences appear to be very great indeed, fact that affects the channel-bed modelling regime. On the Cibin, the lowest monthly averages (1.42-1.63 cum/sec) were recorded in winter (December-March) with a maximum in spring and summer (April, 3.5 cum/sec; July, 10.2 cum/sec). Since comparative data for the Sălişte and the Secaşul Mare are not available, interpreting rates, discharge and channel-bed modelling is a difficult task. So, channel dynamics could be assessed only by field investigations, which nevertheless are not sufficient for establishing correlations. Speaking therefore of the Cibin we would say that the most likely interval for erosion processes to develop is April-June, when high spring waters and floodwaves set in after the snow melts and the rains begin to fall. In summer, high floodwaves are produced only by very intense torrential rains.

The Secaşul Mare (42 km long) drains the Apold Depression from east to west along a distance of 30 km. Its major lefthandside tributaries are the Apold, Dobârca, Gârbova, Pustia, Călnic and Răhău (intermittent or semi-permanent streams). A similar discharge regime have the righthandside tributaries originating from the Amnaş and the Secaş plateaus: the Amnaş, Boz, Draşov, Cut, and Daia. Summer floods have a greater impact on erosion and sediment transport.

The underground waters of the depressions lie in Pannonian and Quaternary deposits of sand and gravel at depths of 30-70 cm in some parts of the Cibin, Sălişte, and Secaşul Mare floodplains and at over 10 m in the southern hills. Unlike the waters in the north of Sibiu which lie on a carbonate bed and

are salty, these underground waters have a very low mineral content (Bălăceanu, 1970).

South-west of Veştem settlement, there is somehow increased water mineralisation due to the Badenian strata; very small surfaces of salt efflorescences are seen on the right and lefthandsides of the Sărata Valley and the Tocile Brook, respectively (Sibiu Depression) and around Miercurea Sibiului Băi (spa resort) in the Apold Depression.

The density of relief fragmentation is an important factor in the dynamics and frequency of modelling processes. It is the resultant of drainage channel density, hence of linear erosion. Without entering into details, we would say that the highest values are registered in the sub-montane hills from the southern flank of the depressions, with differences among them between 3-5 km/sqkm.

The relief energy varies from 80 to 90 m in the hilly zone of the depressions, overlapping areas with frequent geodeclivities of 15°-25°. High values (>25°) are characteristic of cuestas, slide scarps and the steep scarps in the source area of torrential basins. The declivity of semi-horizontal depressionary surfaces is 0°-2°.

The length and aspect of slopes are decisively influencing the overland flow. Moderately long slopes (80 m-150 m up to 250 m in some places) are characteristic of scarps of terraces and cuestas, and of slopes underlain by sandy rocks. On the other hand, structure-controlled surfaces, slopes developed on marl-clays and marl-sands are over 200 m, frequently 250-350 m long. The slopes with a southern aspect, forming the Secaşul Mare cuesta or the lefthandside of the Cerna Vodă, have a high insolation index which accelerates snowmelt, and enhances overland flow, landsliding and gullyng. In the hills extending on the southern flank of the depressions, on slopes with eastem, western and southern aspect, differences are annulled by the vegetation of the forests of Tufari and Mohu (Sibiu Depression); Bercu Roşu and Dumbrava hills (Sălişte Depression) and those covering the Gârbova, Albele and Reciu hills (Apold Depression).

Geological factors. Sedimentary formations are of a Badenian, Sarmatian, Pannonian and Quaternary age; they occur in alternation of impermeable (marls and clays) and permeable strata (sands, loosely-cemented sandstones) with a monocline structure and general SW-WE orientation (Vancea, 1960). The dominance of marl-clays and clay-sands favours landsliding and gullyng.

Vegetation and the anthropic factor. In the Sibiu-Apold depressionary passageway there was a wealth of oak forests (*Quercus robur* and *Quercus petraea*) that used to cover large areas in the past. What has been left of them are a few scores of hectares. There is more forest land in the Sibiu Depression (18%) than in Sălișe (16%) and Apold (5.4%). As the trees were felled, their place was taken by secondary hydro- and mezophile meadows with *Molinia coerula*, *Agrostis stolonifera*, *Alopecurus pratensis*, etc., or by cultures and plantations of a weaker protective value. Pasturelands were degraded by overgrazing (Căplan, Coșcani, Gârbova and the other hills), so that their anti-erosional role has gradually diminished.

Human activity had in the course of time a different impact on the relief. For instance, on the lefthandside of the Cibin Valley there are plantations, terracing (between Veștem and Tălmaciu, Cristian and Orlat), more precisely in the hills closing in the Sibiu Depression in the NW and W), sewerage and embankments (between Șura Mare and Șura Mică, at Șelimbăr, etc.) have all contributed to slowing down erosion.

In the depressionary areas the instability of slopes was augmented by quarries being opened at Gușterița, Bungard, Veștem (Sibiu Depression), Dobârca, Gârbova and Călnic (Apold Depression) or by the exploitation of riverbed gravels and sands. These brought changes in the morphology of channels (e.g. the ballast exploitations in the Cibin Valley, downstream of Orlat, and at Veștem).

Modelling processes and ensuing landforms

In the current evolution of the relief, landslides associated with gullying are the main processes on slope, strongly influencing all the other landscape components. The intensity, rate and diversity of these processes is different in the basins of the Cibin, Săliștea and the Secașul Mare, without, however, reaching the dimensions and frequency of those in the limitrophe Secaș and Hârtibaciu plateaus.

Saying precisely which of these processes played a major role in triggering events is impossible, but there are some areas in which mass movements are obviously mainly involved.

Pluvial denudation and sheet wash phenomena are obvious on steep slopes (over 10°), terrace surfaces or interfluvies. Their intensity depends on declivity, vegetation cover, type of soil and man-changed uses. On low sloping grounds (1°-2°) - the floodplains and terraces of the Cibin, Săliștea and Secașul Mare, used largely as arable land, pluvial denudation disperses the soil particles. In severely

degraded areas (Gușterița and Coșcani Hills, the right valleyside slope of Dobârca and Gârbova) rain drop splash removes and transports the soil particles towards the lower section of slopes, over distances that depend on the intensity and duration of the rainfall.

In the southern hills sheet wash is usually weak-to-moderate, moreover so on forest-protected slopes (e.g. between the Sibiel and Pârâul Vale valleys between the Cerna Vodă and the Mag brooks, and between the Chipeșii and the Călnic valleys) and severer where the vegetation is missing (Căplan, Carpeni, Coșcani hills, etc.) These are lands where effective protection against erosion calls for the management of slopes.

Gullying processes

In general, deep erosion has produced a wide range of forms, and all of them are present in the studied area, from incipient to more evolved ones - rills, ditches, gullies, ravines and torrential organisms in various stages of evolution.

In the Apold Depression, gullying processes are particularly severe on slopes underlain by sands and loose sandstones (Fig. 3). However, they also affect areas subjected to landslides. Covering by far smaller surfaces, gullying is present in the depressions of Sibiu and Săliștea as well. Ravines are usually 5-10 m deep, appearing in the source area in the form of discontinuous small fixed valleys due to piping and down-sagging. Backward erosion of sources and sideways withdrawal of slopes associated with landslides and rock-and-soil falls enlarge the degraded surfaces.



Fig. 3. Righthandside slope of the Călnic Valley subjected to intense gullying (Apold Depression)

Landslides

They are decisively involved in the modelling of slopes in the Sibiu-Apold passageway. Slides may

be superficial (sheet slides) or deep-seated, dislodging materials of 1-2 m or over 2m-thick respectively.

Sheet slides affect much of the area on slopes steeper than 10° (Apold Depression) dominated by marl-clay rocks and a hilly relief. On older, massive fixed slides (steps and monticles - glimee), revealing an older generation, reactivated nuclei appear usually in the median or lower slope section, due to the excess of moisture in the deposits. Attaining a stability threshold takes 1-3 years. In the Apold Depression, such slides are frequently occurring in the basins of the Aciliu, Valea Podului (a righthandside tributary of the Apold Brook), on the righthandside of the Dobârca, Gârbova, Pustia, Călnic, and Răhău valleys and the Secaşul Mare cuesta. The dislodged material often takes the form of small steps or "pseudo-pipings" in some places. They occur either alone or in association.

In general, sheet slides develop in the course of 1-2 months, subsequently their morphology is influenced by other slope modelling processes. The material may travel along one single slide track. Sheet slides may be scores of meters long (80-90 m), 10 m wide and 0.80 - 2.0 m thick. Most scarps have a semicircular or rectilinear form, and are 1.30-2.0 m high.

In the lower section of slopes, they regress backwards, towards the upper slope sector. They are a rich supply source of sediment to the channel. Generally, in the depressions of Sibiu and Sălişte slopes are quite stable. Those affected by sheet slides have varied uses; grazes, arable lands, orchards, and built-in areas. On the righthandside of the Poplaca, Orlat, Cîsnădie valleys and in the hydrographic basins of the Afunde and Fileru (righthandside tributaries of the Cîsnădioara) these types of slope have been mapped.

Deep-seated slides (old and recent) affect small areas in the three depressions, usually slopes developed on marl-clays with intercalations of sands and sand-clays. Sliding occurs both through translation and rotation, dislodging a quantity of 2-5 m-thick materials (over 5 m in some places), 200-350 m long; the well-outlined scarp is 3-10 m-high.

From a morphogenetical viewpoint, slopes are marked by steps, sliding deluvia and monticles (glimee). In the Apold Depression, the best-known glimee-type slides lie NE of Apoldu de Sus settlement in the Potca and Coşcani hills and in the middle sector of the Aciliu and Amnaş basins. In the Sibiu Depression, deep-seated active slides are seen only in the north-eastern part of Sibiu City, more precisely on the SW slope of the Guşterişa Hill.

Investigations conducted so far in various relief units of Romania have shown that Quaternary landslides had played a major role in the modelling of slopes. In the Transylvanian Tableland, pollen

analyses indicate massive sliding during the Boreal, in the Würm or even Subatlantic interphases, the last one being in the very wet Postglacial phase (Morariu et al., 1964). In the Făgăraş Depression, pollen analyses of the peat from the Avrig swamps occurring between the slided heaps on the upper piedmont glacis scarp (Popescu, 1990), date them to the Pleistocene glaciation (Tardyglacial).

Landslides in the studied area fall into two categories: some are old, presumably of an Upper Pleistocene-Holocene age, others belong to the present-day, recent and contemporary periods.

Deep-seated slides represent basic processes in the evolution of the sub-montane hills from the southern flank and the marginal hills from the eastern and northern flanks closing the depressions in. These slides are preserved in the morphology of slopes in the form of steps, sliding deluvia and occasionally monticles (glimee). In general, they have reached an advanced stage of stability and are functionally integrated with the slopes. A second category are the present-day slides, developed in recent history. They represent reactivations of old slides, permanently in a state of imbalance due to heavy precipitation and human activity (Fig. 4).

Geomorphological risk

Geomorphological risk is defined as a probability for the occurrence of some phenomena liable to changing the dynamic balance of slopes, hence with visible effects on the environment. Like changes can be influenced or unleashed by climatic and anthropic factors. The map of geomorphological risk (grounded on present-day geomorphic processes) has a complex character. It makes a synthesis based on repeated investigations of a set of geomorphic factors, highlighting the dynamics of the Sibiu-Apold depressionary passageway relief. The maps used to bring it as close to reality as possible were: map of morphostructure and morpholithology; map of land use, map of slopes, and map of slope aspect. This map of geomorphological risk is an important achievement, because: 1. It shows the degradation of unstable or fragile environments; 2. It facilitates the zonation of areas fit for construction works; 3. It defines and assesses the constraints or limitations posed by natural landforms. The categories of geomorphological risk in the Sibiu-Apold depressionary passageway are grouped by degree of stability and dynamics of moving material, intensity and frequency of landslides associated with gullyng and human impact. The five categories of geomorphological risk have been listed under three classes: high, moderate and low (Fig. 5).

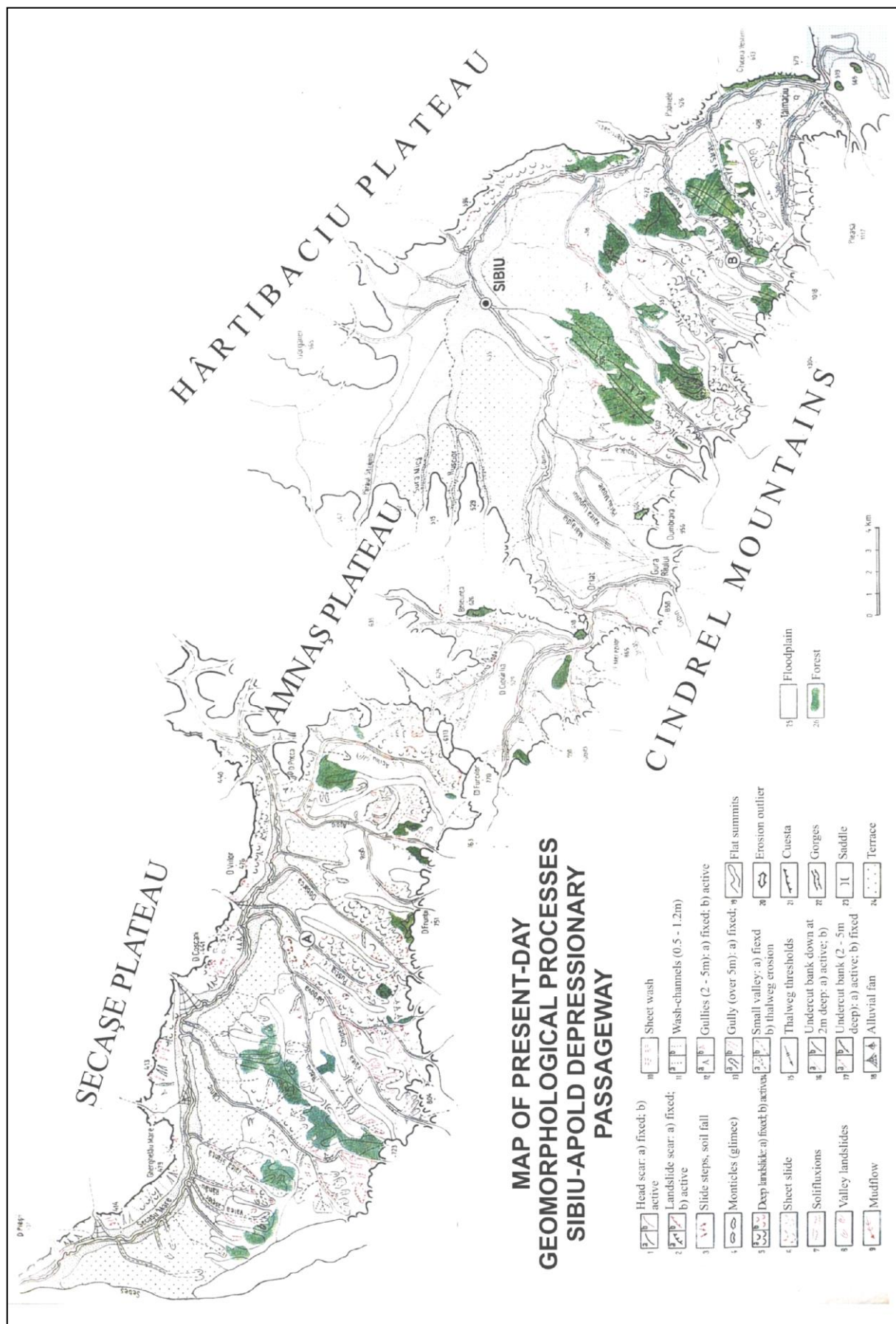


Fig. 4. Map of present-day geomorphological processes. Sibiu-Apold depressionary passageway

High geomorphological risk

Slopes at high geomorphological risk have a big morphodynamic potential, 25-35° declivity, 80-150 relief energy, dominantly south and east aspect and a frequency of 22-30 watercourses/sqkm. In general, these slopes are covered with vulnerable unproductive soils and erosion-control works are suggested. High risk areas represent 17.4% in the Apold Depression, 3.5% in the Sibiu Depression being totally absent in the Săliște Depression. This risk category includes two situations;

- *slopes of high risk from deep-seated slides often associated with gullyng.* In the Apold Depression: the Aciliu Basin, the western scarp of Amnaș Plateau, Dobârca and Gârbova hills, Viilor and Coșcani hills (Secașul Mare cuesta). The intensity and frequency of landslides and the backwards regression of the Aciliu and the Amnaș sources maintain a permanent imbalance on slope, being a threat to the railway and the national highway between Aciliu-Apoldu de Sus settlements, corresponding to the unstable sector of the Mureș-Olt watershed.

Despite marked slope instability, works of consolidation, draining, etc. proved to be efficient. In the Sibiu Depression: the Gușterița Hill (NE of Sibiu City).

- *slopes at high risk from gullyng associated with severe sheet wash* are found in the Gârbova Hill, the lefthandside slope of the Reciu and the Călnic valleys (Apold Depression) and the righthandside slope of the Orlat Valley, as well as in the Fântânele Basin, east of Rășinari locality.

Moderate geomorphological risk

The slopes in this category have 15-20° declivity, eastern and southern aspect, 50-80 m relief energy, frequency of watercourses 15-20 sqkm. Stability is reduced due sheet and deep-seated slides. This category of slopes represents 20% in the Apold Depression, 22.5% in the Săliște Depression and 16.6% in the Sibiu Depression. Two situations occur in terms of the frequency and intensity of each process:

- *slopes at moderate risk due to sheet slides associated with gullyng* in the Apold Depression: the righthandside slope of the Amnaș Valley, the upper Pustia Basin and the Călnic Hill; the Săliște Depression: the lefthandside slope of the Cerna Vodă and the Mag valleys; the Sibiu Depression: part of the western scarp of the Hârtibaciu Plateau (closing in the depression to the east, between Bungard settlement and the Hârtibaciu Valley);

- slopes at moderate risk due to gullyng and periodical reactivation of sliding occur in the Ghergheleu and Călnic hills and on the lefthandside slope of the Reciu Valley (Apold Depression), a smaller area on the righthandside slope of the Călnic Valley, and the middle basin of the Sărata (Sibiu Depression).

Low geomorphological risk

This category of slopes has the lowest morphodynamic potential; 2-15° declivity, frequency of watercourses 8-10 sqkm. Sheet wash is weak and reactivation of some gully and ravine sources is ephemeral. These slopes represent 23% in the Apold Depression, 53.7% in the Săliște Depression and 26.4% in the Sibiu Depression. Landforms are in an advanced stage of evolution due to mass movements and gullyng. Most of them are fixed by vegetation.

Slopes falling into this category are: both sides of the Tocilelor Brook, the righthandside of the Sadul and of the Sibiel (Sibiu Depression); small areas on the lefthandside of the Dobârca, Gârbova and Reciu valleys downstream of the homonymous settlements, and Răhău (Apold Depression).

Much of the Sibiu-Apold depressionary passageway is *risk-free* in the lower parts of the depressions: floodplains and terraces with stable grounds, declivity around 2° (excepting the fixed terrace scarps: 5-15°), relief energy 5-10 m. Risk-free areas cover 39.6% in the Apold Depression, 23.8% in the Săliște Depression and 53.3% in the Sibiu Depression. The map of geomorphological risk also presents channels at moderate risk of overflowing during floodwaves: the Secașul Mare and its lefthandside tributaries - the Aciliu, Apold, Dobârca and Călnic (Apold Depression); the Săliște and its tributaries - the Sibiel, Pârâul Vale and Cerna Vodă (Săliște Depression); the Seviș, Cislădie and Sadu (Sibiu Depression) (Fig. 5).

At moderate risk from overflowing is the Cibin between Veștem and Tălmăciu settlements. An important role in preventing and diminishing floodwaves on the Cibin has Gura Râului reservoir (on the river) commissioned in 1980, as well as the draining system that covers 2,300 hectares between the settlements of Șura Mare and Ruscior. It would be advisable to have a similar draining system in the Secașul Mare floodplain (Cunța-Cut segment) where the underground water sheet lies close to the surface (-0.4-0.5 m) and the thin stratum of clay favours water stagnation. Draining would also eliminate gleization of the soil.

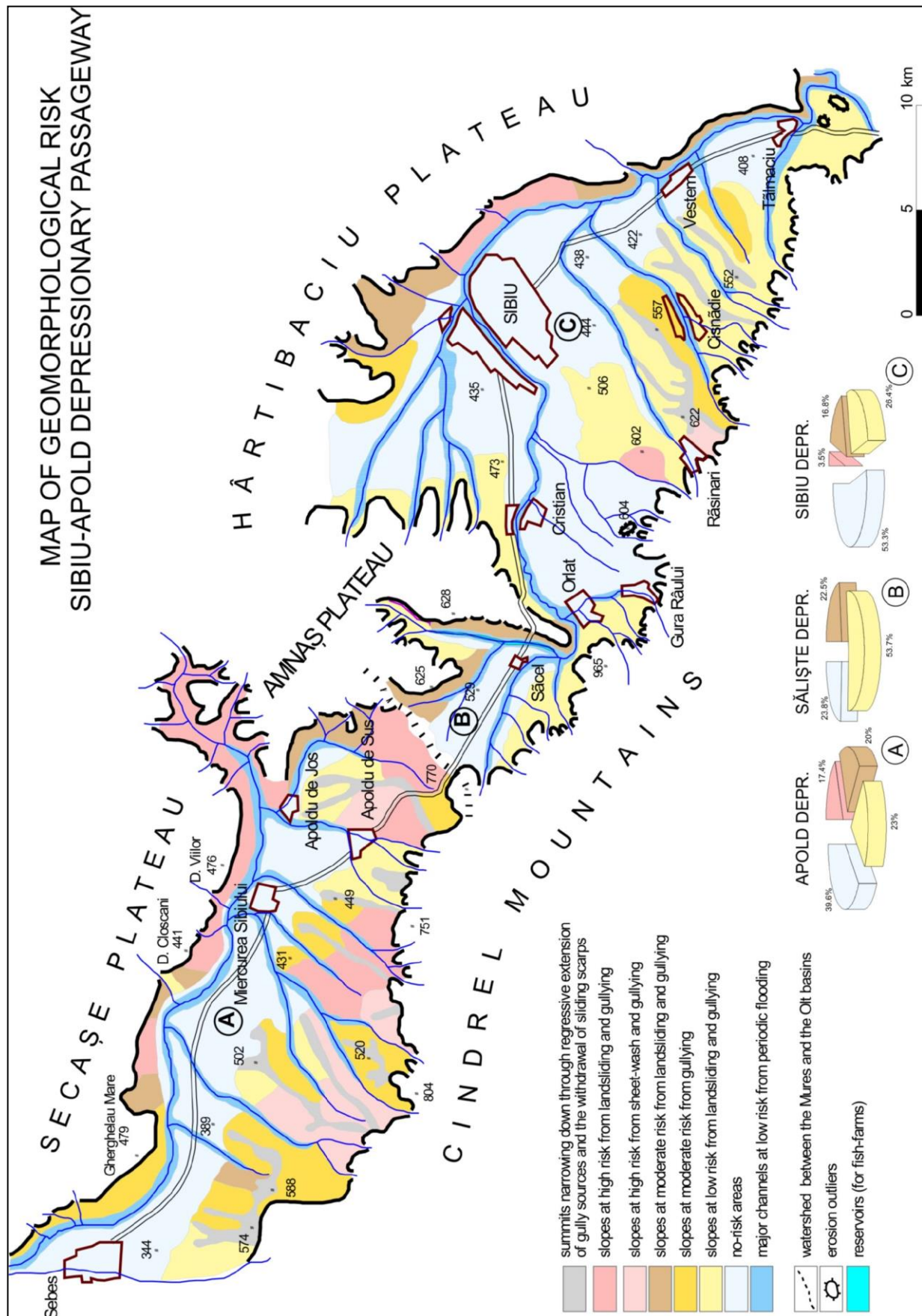


Fig. 5. Map of geomorphological risk. Sibiu-Apold depressionary passageway

Conclusions

The present-day modelling of the study area is a complex process begun in the Pleistocene and continued through the Quaternary to the present. Modelling processes are not isolated, they are associated with similar or different categories of processes. Slopes, for example, undergo several types of mass movements and erosion processes, differing in space or time. It is fairly difficult to say precisely which of these processes have been fundamentally involved in the modelling of slopes. However, in the present stage of relief evolution landslides associated with gullying are the main processes that maintain a permanent state of

imbalance on slope. In general, they are triggered by extreme precipitation and human activity.

Given the specific morphodynamic evolution, as well as the intensity and extension of imbalanced areas, a map of geomorphological risk, based on some characteristic parameters, has been drawn up, with highlight on high, moderate and low risk zones. They relate to the graded complexity of the slope processes studied and the affected area (together with its structure-controlled substrate).

Taking effective measures for a more appropriate use of the land (by observing its characteristic features), simultaneously with improving and protecting the environment, means conducting further investigations by methods and techniques which, so far, are not available to us.

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Response of Soil Erosion to Land Use Change with Particular Reference to the Last 200 Years (Julian Alps, Western Slovenia)*

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Key Words: geomorphology, erosion, erosion modelling, Gavrilović equation, land use, land use changes, Alps, Slovenia.

Abstract. Erosion changes in the Upper Soča Valley (Julian Alps, western Slovenia) during the last two centuries are discussed in the article. Erosion was calculated using a model in which land use was one of the principal factors. The empirical Gavrilović equation that has been partly modified by Lazarević was used. Erosion calculations covered five years: 1827, 1896, 1953, 1979 and 1999, enabling us to establish changes in the erosion process due to a change in land use over the last 200 years. The total annual erosion in the study area amounted to 4.76 million m³ in 1827, and it was approximately the same also in 1999. Within that interval erosion increased, reaching 5.72 million m³ in 1953, with a presumed maximum during the 1920's. In 1827, specific erosion was 133.4 tons/hectare/year, in 1999 it was 135.5 tons/hectare/year, and in the meantime, in 1953, it reached 160.3 tons/hectare/year. The study shows that erosion changes may be established also by means of historical sources of land use, if a proper model is employed.

1. Introduction

Soil erosion and sediment deposition processes are determined by four main factors: regolith type, climate, topography and land use. Although all of these factors interact with human activity to a certain extent, land use is the most 'manageable' factor (Van Rompaey et al., 2003). Changes in soil erosion in the western part of the Julian Alps in Slovenia (591.5 km²), more precisely in the drainage basin of the Soča river (Fig. 2), during the last two centuries were established exactly through the land use changes.

Required for such calculation was the erosion model in which land use was employed as one of the principal factors for the calculation of soil erosion. Selected was the empirical Gavrilović equation (Gavrilović, 1962; 1970; 1972) partly modified by Lazarević (1968a; 1985). The equation is applicable to the analytical determination of erosion coefficients and the quantification of erosion and average annual sediment yield. This method has been widely used in Slovenia and in the western Balkans for the last 30 years to predict erosion processes and implement torrent regulation and other erosion control works (Gavrilović, 1988; Globevnik et al., 2003; Zorn, 2008).

The study of Klanghofer, Hintersteiner and Summer (2002) has shown that the expansion of farm land into hilly/mountainous areas with steep slopes and changes in land use as well as in land management techniques significantly increased the soil erosion rates as well as the sediment yield from different subcatchments within the Danube drainage basin (in Austria) in the 1950–1990 period.

The trend was different in the Slovenian Alps, where an extensive process of depopulation and abandoning of agrarian activities took place, which led to the decline in farming areas and increase in the percentage of forests (Petek, 2005a; 2005b). Erosion played a greater role about a century ago than it does today, because the proportion of cultivated areas was substantially larger. But at the beginning of the 19th century erosion is supposed to have been approximately on the same level as at present. The amount of eroded material has decreased in the last decades, due to natural, social and economic factors. Reforestation is the key process. In the past decades different authors supposed (for sources see Komac, Zorn, 2005; 2007; Hrvatin et al., 2006) that annual soil erosion in Slovenia amounts to 5 to 6 millions m³ of

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material (~2.5 to ~3.0 m³/ha). But according to the data that are presented in this article and according to the measurements of erosion in the Slovene Istria region (Zorn 2008), we can conclude that, in the past, soil erosion has been almost certainly underestimated.

2. Methods

2.1. Gavrilović's method for the calculation of annual soil erosion

For a quick and easy quantification of soil erosion quite a number of erosion models have been developed, the majority of which were based on the combination of short-term measurements and mathematical formulas. These are the so-called empirical models. Thus, the empirical models are based on the link-up of a dependent variable with a set of measured or assessed independent variables by means of regression analysis (Staut, 2004). Of such models, most widely used is the American erosion model USLE (Universal Soil Loss Equation), developed by Wischmeier and Smith (1965); based on it are the contemporary most widely spread empirical models (e.g. RUSLE).

In the western Balkans the most widely used have been the so-called Gavrilović equation (1962; 1970; 1972) and its modifications (Lazarević, 1968a; 1985; Pintar, Mikoš, Verbovšek, 1986), which are similar to the USLE model (Mikoš, Fazarinc, Ribičič, 2006). The advantage of this

equation, according to Z. Gavrilović (1994), lies in the fact that it is not meant primarily for the calculation of soil erosion on arable lands – for this purpose the USLE method was worked out – but was developed for the hydro-regulation needs and is suitable for the calculation of soil erosion irrespective of land use. The equation is based on several years of measurements performed on several tens of erosion plots in central Serbia (Lazarević, 1968b).

S. Gavrilović (1962; 1970; 1972) proposed an analytical equation for determining the annual volume of detached soil due to surface erosion:

$$W_a = T * P_a * \pi * \sqrt{Z^3} * F$$

where:

W_a = total annual erosion (m³/year)

T = temperature coefficient

P_a = average yearly precipitation (mm)

Z = erosion coefficient

F = study area (km²).

$$T = \sqrt{\frac{T_0}{10}} * 0.1$$

T_0 = average yearly temperature (°C)

$$Z = Y * Xa * \left(\varphi + \sqrt{J_a} \right)$$

Y = soil erodibility coefficient

Xa = soil protection coefficient

φ = erosion and stream network development coefficient

J_a = average slope of the study area (%).

Table 1. Descriptive factors used in the Gavrilović model (based on Lazarević, 1985; Vente, Poesen, 2005)

Soil protection coefficient	Xa
Mixed and dense forest	0.05–0.20
Thin forest with grove	0.05–0.20
Coniferous forest with little grove, scarce bushes, bushy prairie	0.20–0.40
Damaged forest and bushes, pasture	0.40–0.60
Damaged pasture and cultivated land	0.60–0.80
Areas without vegetal cover	0.80–1.00
Soil erodibility coefficient	Y
Hard rock, erosion resistant	0.1–0.3
Rock with moderate erosion resistance	0.3–0.5
Weak rock, schistose, stabilised	0.5–0.6
Sediments, moraines, clay and other rock with little resistance	0.6–0.8
Fine sediments and soils without erosion resistance	0.8–1.0
Erosion and stream network development coefficient	φ
Little erosion on watershed	0.1–0.2
Erosion in waterways on 20–50% of the catchment area	0.3–0.5
Erosion in rivers, gullies and alluvial deposits, karstic erosion	0.6–0.7
50–80% of catchment area affected by surface erosion and landslides	0.8–0.9
Whole watershed affected by erosion	1.0

For details on the coefficients used in the article (Table 1) see S. Gavrilović (1962; 1970), Z. Gavrilović (1988), Lazarević (1968a; 1985), Staut (2004) or Vente and Poesen (2005).

The great generality of variables and parameters used in the Gavrilović method renders the assessment of erosion processes possible also in the case of rough information that can be obtained even without fieldwork. Thus, the insight into the past erosion activity is possible as long as good historical (cartographic) bases and aerial shots exist, which can be integrated into the national network of coordinates (Petek, Fridl, 2004), as well as climatic data strings (Staut, 2004).

Calculations following the Gavrilović equation are done by hydrogeographical units, i.e. by drainage basins or by sections of drainage basins. Our calculations had to be done by cadastral municipalities, the boundaries of which correspond pretty well to the boundaries of drainage basin sections. Thus, the cadastral municipalities could be assumed as the hydrogeographical units.

2.2. Data preparation and GIS manipulation techniques

Erosion was calculated for five terms, i.e. for the years 1827, 1896, 1953, 1979 and 1999. On the basis of the obtained calculations for these terms, it was possible to establish the changes in the erosion process as a consequence of land use changes in the

last 200 years (Komac, Zorn, Gabrovec, 2007; Zorn, Komac 2008).

The data on land use for the year 1827 were obtained on the basis of the Franziscan cadastre, i.e. the land cadastre measurement project that was started according to the emperor's patent in 1817 (the territory of Slovenia was part of the Austrian empire at that time), and completed for the majority of Slovenian territory by the year 1828 (Petek, Urbanc, 2004). The data were taken from the written part of the Franziscan cadastre (Fig. 1), separately for each cadastral municipality (Petek, 2005a).

From the land cadastre the data on land use for the year 1896 were also taken (Petek, 2005a). In order to establish land use changes after the Second World War, we likewise used the data from the land cadastre, aggregated on the level of cadastral municipalities for the years 1953, 1979 and 1999 (Petek, 2005b). Calculations were done in the form of raster with the basic cell size of 25 x 25 metres, by means of the *Idrisi* program package.

The data on the lithological structure of the area were obtained from a digital geological map at a scale of 1:100,000 (Buser, 1986; Jurkovšek, 1987) and the data on soils from a digital pedological map at the scale of 1:25,000 (Pedološka, 1999), while the data on landforms were obtained from the digital elevation model 25 x 25. Since no data on climate are available for older periods, the data set for the 1961–1990 reference period was applied to the entire period (Zupančič, 1995).

Fig. 1. Formular from the Franziscan cadastre for the registration of the size of individual land use categories by cadastral municipalities (source: Archives of the Republic of Slovenia)

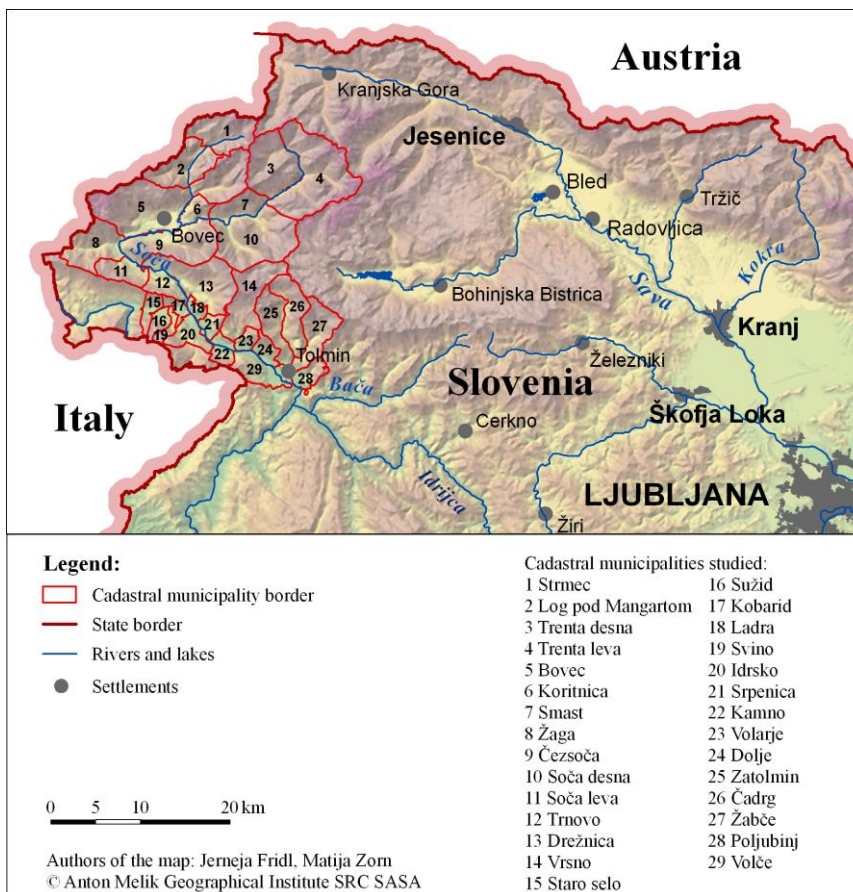


Fig. 2. Map of the study area

4. Results

Between 1827 and 1900 changes in land use in the Slovenian Alps were minimal, as they were registered on less than half a percent of the surface. Among the processes involved in the change of land use, intensification prevailed. Intensification primarily involved the transformation of grassland into cultivated fields. Between 1900 and 1953, selective limitation of cultivated fields only to the most favourable land began. Abandoned fields turned into grassland (primarily meadows), and thus overgrowing with grass dominated during this period, when varying social conditions had the greatest influence on the changes in land use. Between 1953 and 2000, changes in land use were the greatest, with the prevailing afforestation. However, we established a difference in the degree of afforestation between the first and the second half of this period: in the first half, changes in land use were generally the most intensive in the entire period studied. Between 1953 and 1979, the largest proportion of agricultural land was transformed to forests. Between 1979 and 2000, the intensity of afforestation decreased slightly, possibly due to the

legislative changes and measures adopted after 1970 to stimulate agriculture. Between 1900 and 2000, a quarter of the land in Slovenia's alpine region was subject to afforestation that took place mostly because of economic changes (Petek, 2005b).

The total annual erosion in the discussed area amounted to 4.76 million m³ in 1827, and it was approximately the same also in 1999. In the meantime erosion increased and reached 5.72 million m³ in 1953, with presumed maximum during the 1920's. The main cause of the increase in erosion was the agrarian overpopulation of the area, which gave rise to economic emigration towards the end of the 19th century.

Interesting are the specific values of erosion which point to the great significance of erosion focal points. These are numerous in the studied area, especially on the steep dolomite slopes, scree slopes, steep slopes in quaternary sediments and partly also on the flysch slopes dissected by a dense network of valleys. In 1827, the specific erosion amounted to 133.4 tons/hectare/year, in 1999 it was 135.5 tons/hectare/year, and in the meantime, in 1953, it reached 160.3 tons/hectare/year (Fig. 3; Fig 6).

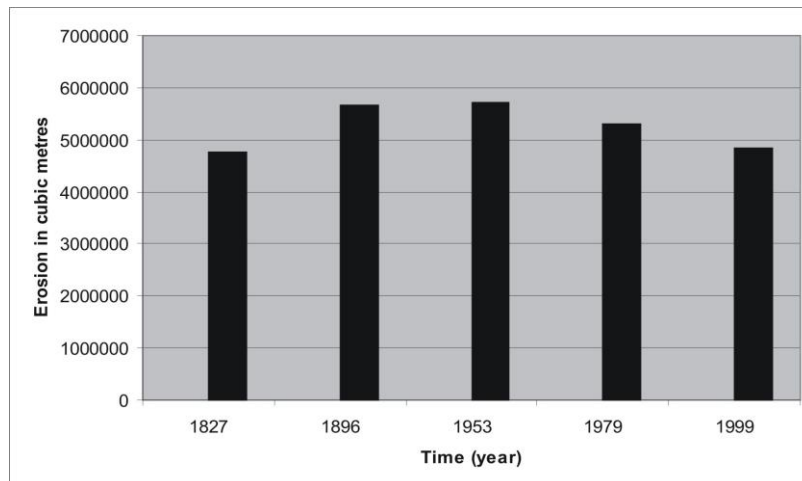


Fig. 3. Erosion in the Upper Soča region between 1827 and 1999 (in m³ per year)



Fig. 4. The Svinjak mountain near Bovec (Upper Soča valley) in the beginning of the 20th century (left; photograph: M. Klodič) and in the beginning of the 21st century (right; photograph: Miha Pavšek)

The causes for such a course of development can be assumed to be found in land use changes, which reflect the complex socio-economic as well as the natural processes which were going on in the discussed area during the past two centuries.

The percentage of arable fields decreased throughout the region, and declined by a half between the years 1827 and 1999. Also, the percentage of pastures declined by about 40% and the percentage of meadows declined by 15%. In the same period the percentage of forest areas increased by one third (Table 2; Fig. 4). The protective role of ever larger forest areas on steep slopes significantly contributed to the reduction of erosion in the last decades. The decline in percentage of pasture areas, which are, due to high altitude and higher inclinations, more intensely exposed to erosion than meadows, also had a significant impact. The

decrease in arable field areas had but relatively little influence on the erosion reduction. Arable fields on the discussed territory lie only at the bottom of valleys or on (river) terraces, where erosion is less due to slight inclinations of the surface.

With regard to the area occupied by a certain land use category, it can be concluded that the increased erosion in the past century was correlated with the insufficient percentage of forests and the high percentage of pasture areas. The extent of alpine pastures began to decrease only after the economic transition of the 1970s, when the active population got employed more intensely in industry and its survival no longer depended on animal husbandry or pasturing on the Alps or even haymaking on steep mountain meadows. The reduction of erosion in the past decades also resulted from a smaller percentage of fields (arable areas).

Table 2. Percentage of areas by land use categories in the years 1827, 1896, 1953, 1979 and 1999 (Petek, 2005b)

%	1827	1896	1953	1979	1999	Difference 1827-1999	index
Field	7.73	7.93	7.37	4.77	4.04	-3.68	52.31
Meadow	20.74	21.30	20.81	20.61	17.96	-2.79	86.57
Pasture	27.77	23.43	21.73	18.72	16.09	-11.68	57.95
Forest	35.98	36.50	37.84	42.86	47.77	11.79	132.77
Other	7.78	10.84	12.25	13.03	14.14	6.36	181.70

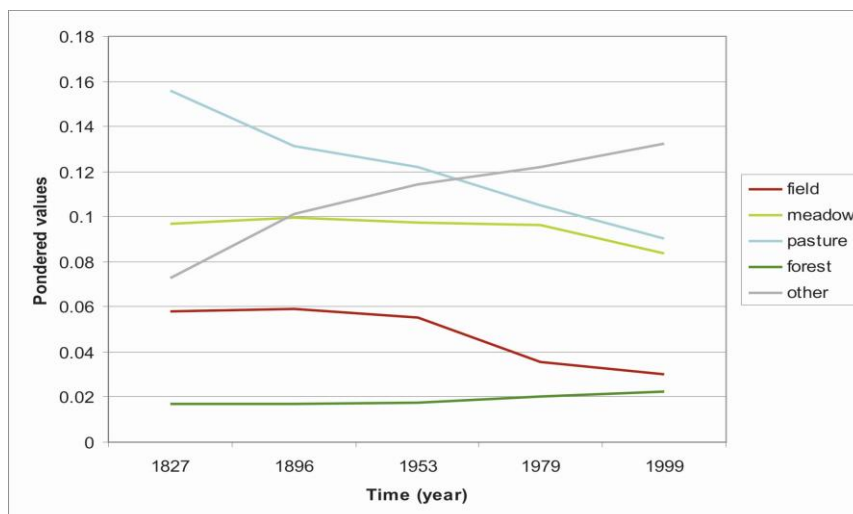


Fig. 5. Pondered erosion values as to erodibility coefficient in the years 1827, 1896, 1953, 1979 and 1999

In order to compare more easily the erodibility of lands of different land use categories, we standardized the areas. In this way it has been established that the pastures are the ones that exert the relatively most intense influence on the erodibility of a certain area. Also the influence of mountain pastures and meadows is very strong in the discussed mountainous region. It is also evident from Figure 5 that the erodibility of lands during the past times mainly depended on the decrease in arable fields and pasture areas and the increase in the extent of so called other areas, including bare land and urbanised areas.

Although we have established that the results faithfully mirror the socio-economic development of the area, we have to be cautious about the source used. In the recent years, the land cadastre of Slovenia has no longer been a very reliable source of information on the actual land use, but only a useful source of information on the ownership situation. The land use map, updated and edited by the Ministry of Agriculture, Forestry and Food of the Republic of Slovenia (Raba, 2002), is a more relevant source. Judging from this map, the total

annual erosion in 2000, for example, was lower (3.05 million m³) than that of 1999 (4.84 million m³), calculated from cadastral data.

However, it should be noted that the above-mentioned data on erosion represent the average, calculated by means of the model. To be sure, extreme events are not taken into account in such calculations (Staut, 2004); the amount of released material during such events equals the size class of the total annual erosion in the Upper Soča region. Thus, at the village of Log pod Mangartom, the debris flow alone in the year 2000 transported over a million of m³ of material to lower-lying locations (Komac, Zorn, 2007). Therefore, the original Gavrilović equation was modified in Slovenia, so that instead of the annual precipitation amount the data on maximum daily precipitation were employed (Pintar, Mikoš, Verbovšek, 1986).

According to the calculation of Mikoš, Fazarinc and Ribičič (2006) average annual sediment production in the discussed area amounts to about 2200 t/km² or about 1400 m³/km². Average annual sediment load is about 200 m³/km², and average delivery ratio 0.31. The figures are rather low

comparing to the results of our analysis for annual sediment production (8047 to 9670 m³/km²). But taking into account the fact that earthquakes and heavy rainfalls are rather frequent in the Upper Soča River basin the figures become more realistic.

Earthquake-induced rockfalls and rainfall-induced landslides may release sediment in excess – about 125,000 m³/km² annually (Mikoš, Fazarinc, Ribičič, 2006) i.e. about twelve times higher comparing to average sediment production in the area.

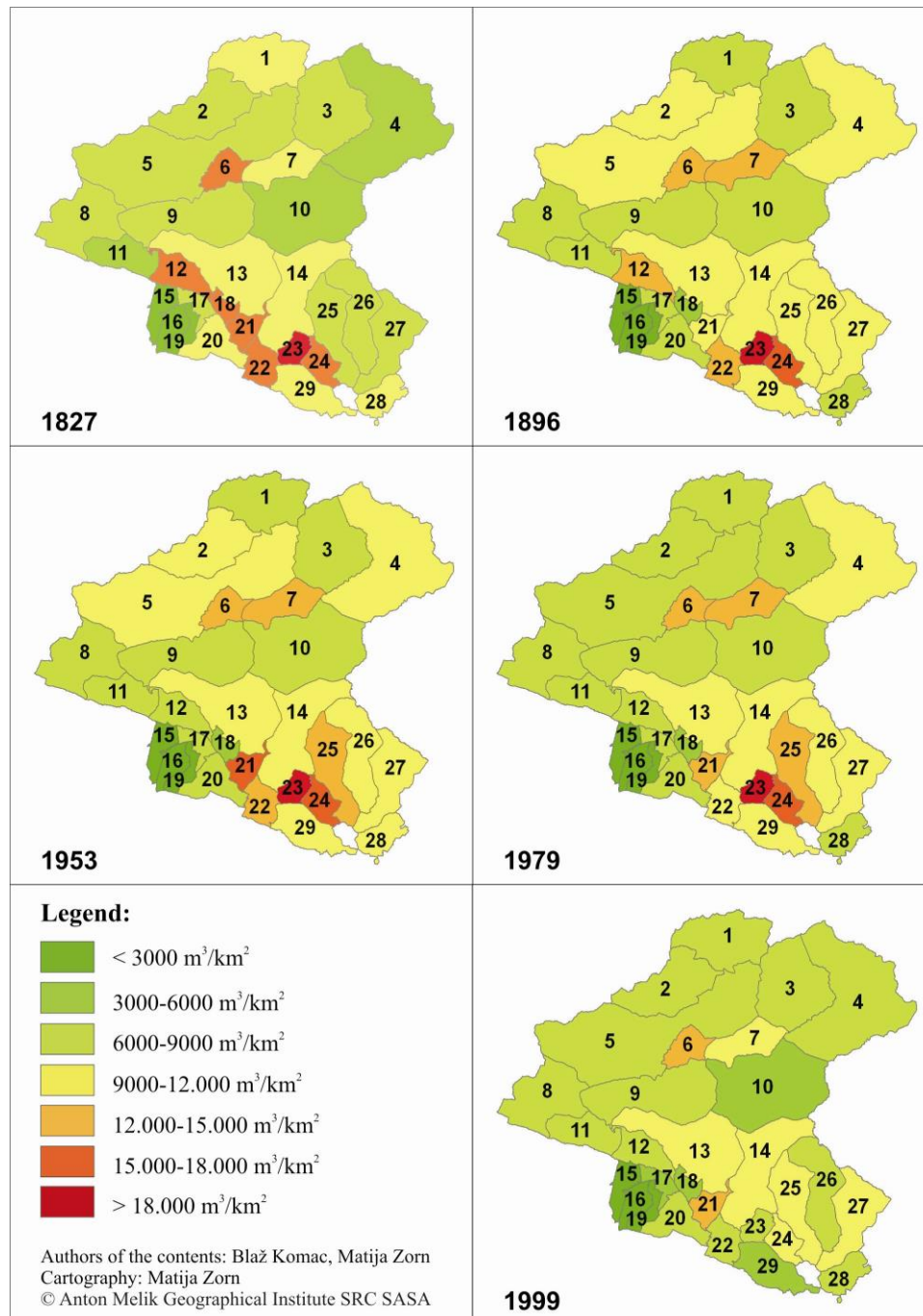


Fig. 6. Specific annual sediment production according to cadastral municipalities in the studied years (for numbers of cadastral municipalities see Figure 2).

5. Conclusion

The paper shows that historical sources on land use are very useful for establishing the changes in erosion, if a proper model is employed. Geomorphic

response to land use is non-linear: a small change in the percentage of arable land usually results in relatively big changes in erosion risk and sediment delivery (Van Rompaey et al., 2003). Thus, an increase in the percentage of arable land results in a

faster than linear increase of the mean soil erosion rate in the drainage basin, because the slope gradients of the newly deforested areas are systematically higher than the slope gradients of the original arable land. Afforestation, on the other hand, results in faster than linear decrease of the mean annual soil erosion rate in the drainage basin, because parcels on steep slopes are more likely to be converted into forest. For example, the decrease in percentage of arable lands in the drainage basin by 5%, may result in a 8.5% lowering of the mean annual soil erosion rate and as much as 13.5% lowering of sediment yield (Van Rompaey et al., 2003).

Complex studies of recent morphogenetic processes together with the comprehension of climate changes and human interventions in the landscape are necessary to understand the relationship between past land use changes and erosion. Human induced processes, such as land use

changes, usually have long-term influences on landscape evolution. The changes in erosion processes, related to human-geographical processes, are thus more a distant reflection of past processes than a consequence of recent anthropogenic processes. In our case the loosened natural balance from previous phases of landscape evolution, especially to the climax of pasture usage in the 2nd half of the 19th century lead to bareness of soil and increased erosion in the 1st half of the 20th century. Even though the landscape has been renewing since then, the natural but labile equilibrium may suddenly collapse, leading to intensive erosion processes, such as rill erosion, strong denudation and slope processes, especially landslides. One of the most difficult tasks of geomorphology is to properly comprehend and evaluate weights of influence factors, e.g. natural and anthropogenic processes in the landscape with different and overlapping evolution periods.

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Analyse du paysage géomorphologique des Hédil (Tunisie) par MNT

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Mots clés: SIG, Modèle Numérique de Terrain, Analyse morphostructurale, Discontinuités.

Résumé. Cet article présente l'intérêt de l'utilisation des Systèmes d'Information géographique (SIG) en géomorphologie et, plus particulièrement, pour l'analyse morphostructurale. En effet, le modèle numérique de terrain (M.N.T), et les cartes qui en découlent, offrent de larges possibilités pour la reconnaissance du paysage géomorphologique et notamment pour l'analyse morphostructurale du terrain à travers, entre autres, la détection de discontinuités topographiques (ruptures de pentes). Une base de données géoréférencées **Hedil-SIG** a été constituée sur la région des Hédil (Nord de la Tunisie), intégrant les paramètres clefs expliquant la géomorphologie structurale de la région. L'analyse multisources (MNT, cartes dérivées, géologie, hydrographie) montre que la région des Hédil est caractérisée d'une part, par un relief fortement accidenté avec une structuration complexe et un linéament multidirectionnel à dominante SW – NE et, d'autre part, par une importante dynamique actuelle résultant de l'activité des eaux courantes et d'une topographie très marquée par l'empreinte des données structurales (lithologie et tectonique).

Key words: GIS, Digital Elevation Model, morphostructural analysis, Discontinuities.

Abstract. This article presents the interest of the utilization of Geographical Information System (GIS) in geomorphology and, more particularly, for the morphostructural analysis. Indeed, cards derived of the Digital Elevation Model (DEM) offer wide possibilities for the recognition of the geomorphological landscape and notably for the morphostructural analysis of the terrain through, among others, the topographic discontinuity detection (gaps of slopes). A georeferenced database has been constituted, named Hedil-SIG on the region of Hédil (North of Tunisia), integrating key parameters explaining the morphostructural analysis of the region. The analysis multisources (DEM, cards derived, geology data, hydrography) show that the region of Hédil is characterized on the one hand, by a relief strongly injured with a complex structuring and a multidirectional outline to dominant SW-NE and, on the other hand, by an important current dynamics resulting the activity of waters and a topography very marked by the structural data imprint (lithology and tectonic).

La région des Hédil et ses environs, en Tunisie septentrionale (Fig. 1), s'étendent sur une superficie totale d'environ 640 km², l'altitude moyenne de la région est de 300 m seulement.

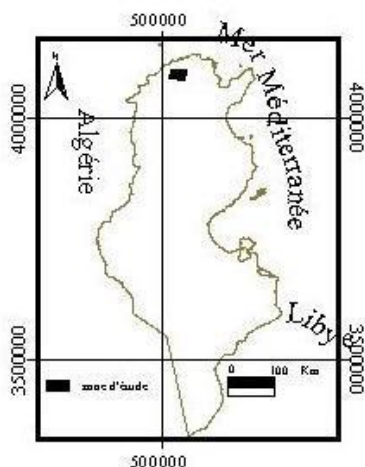


Fig. 1. Localisation de la zone d'étude

Malgré la faiblesse de ses altitudes et de sa faible extension, cette région est très accidentée. Le nord de la Tunisie présente une assez grande variété de paysage. Celui des Hédil est une zone charnière se situe entre deux ensembles bien distincts : le domaine des écaïles au sud-est et le domaine de nappes de charriage au nord-ouest.

Le contact entre ces deux domaines se fait par le complexe salifère du trias dont la mise en place reste un sujet de controverse entre les différents auteurs à la fois géologues et géomorphologues (Kujawski H., 1969; Miossec A., 1979). Dans ce contexte, nous avons essayé de voir si les SIG pouvaient apporter une information supplémentaire à la compréhension de ce paysage géomorphologique en s'appuyant sur une analyse morpho-structurale (Collina-Girard J.1989).

Dans cet objectif, nous avons réalisé un Modèle Numérique de Terrain (MNT) dans la région des Hédil. Ce MNT nous a servi à réaliser des cartes

dérivées (cartes de pentes, cartes d'ensoleillement, carte des expositions, carte des classes d'altitudes). Ces dernières, croisées à d'autres couches d'information telles que la géologie et l'hydrographie, nous ont permis une meilleure appréhension du paysage de géomorphologie structurale de la région des Hédil. L'ensemble a été intégré dans une base de données géoréférencées et multi-thèmes (Rabia M.C., 1998) **Hedil-SIG** et pilotée par le logiciel Arc-Info et Arc-View.

Deux types de données cartographiques sont intégrés au SIG sous forme de courbes, points et polygones :

- la carte géologique au 1/50000 réalisée par le Service Géologique de Tunis,
- les quatre cartes topographiques au 1/25000, correspondant au découpage de la carte au 1/50000^e des Hédil n°11 de l'Office de Topographie et de Cartographie de Tunis.

Les données ont été numérisées à l'aide d'une table à digitaliser A0 pilotée par le logiciel Arc-Info sous PC. Les premiers traitements ont donné lieu aux couches d'information suivantes :

- l'élaboration du MNT de la topographie ;
- les cartes dérivées :
 - o la carte de l'hypsométrie,
 - o la carte des pentes,
 - o la carte des expositions,
 - o la carte d'ensoleillement.

A cela se sont ajoutés les résultats d'analyse statistique, à savoir :

- o le calcul des rosaces de direction des failles et celui du réseau hydrographique en terme de fréquence et de longueur,
- o la distribution spatiale des failles et des cours d'eau.

Méthodologie

Nous avons numérisé systématiquement toutes les courbes de niveau et les points cotés qui couvrent la région à partir des quatre cartes topographiques au 1/25000 de la région d'étude. Dans l'objectif d'obtenir un MNT très proche de la "réalité", nous avons aussi numérisé le réseau hydrographique pour l'intégrer dans les paramètres de la réalisation du MNT. Nous avons aussi numérisé les données de la géologie (le tracé des failles et les affleurements).

A – Le Modèle Numérique de Terrain

L'élaboration du MNT repose sur la méthode d'interpolation triangulaire (TIN) de Delaunay

(Bjorke J.T. et al., 1993). Cette méthode est basée sur l'idée qui consiste à n'utiliser que les points voisins du nœud considéré. En effet, trois points seulement sont utilisés pour définir un maillage triangulaire. Les points de mesures sont les sommets du triangle. Des interpolations linéaires peuvent aussi être calculées à l'intérieur de chaque facette triangulaire. Cette interpolation sera d'autant plus réaliste que les formes des facettes se rapprochent du triangle équilatéral. L'intégration du réseau hydrographique vient renforcer le "réalisme" du modèle numérique de terrain en imposant à cette méthode d'interpolation la prise en compte des talwegs comme des limites naturelles dans la délimitation des triangles limitrophes des cours d'eau. Les effets de "facette" sont dans un second temps atténués par un lissage pour produire une matrice de cellules à taille égale appelée grid ou lattice. L'unité surfacique élémentaire du MNT est de 25 m correspondant à 1mm sur la carte topographique au 1/25 000e.

B - Les cartes dérivées :

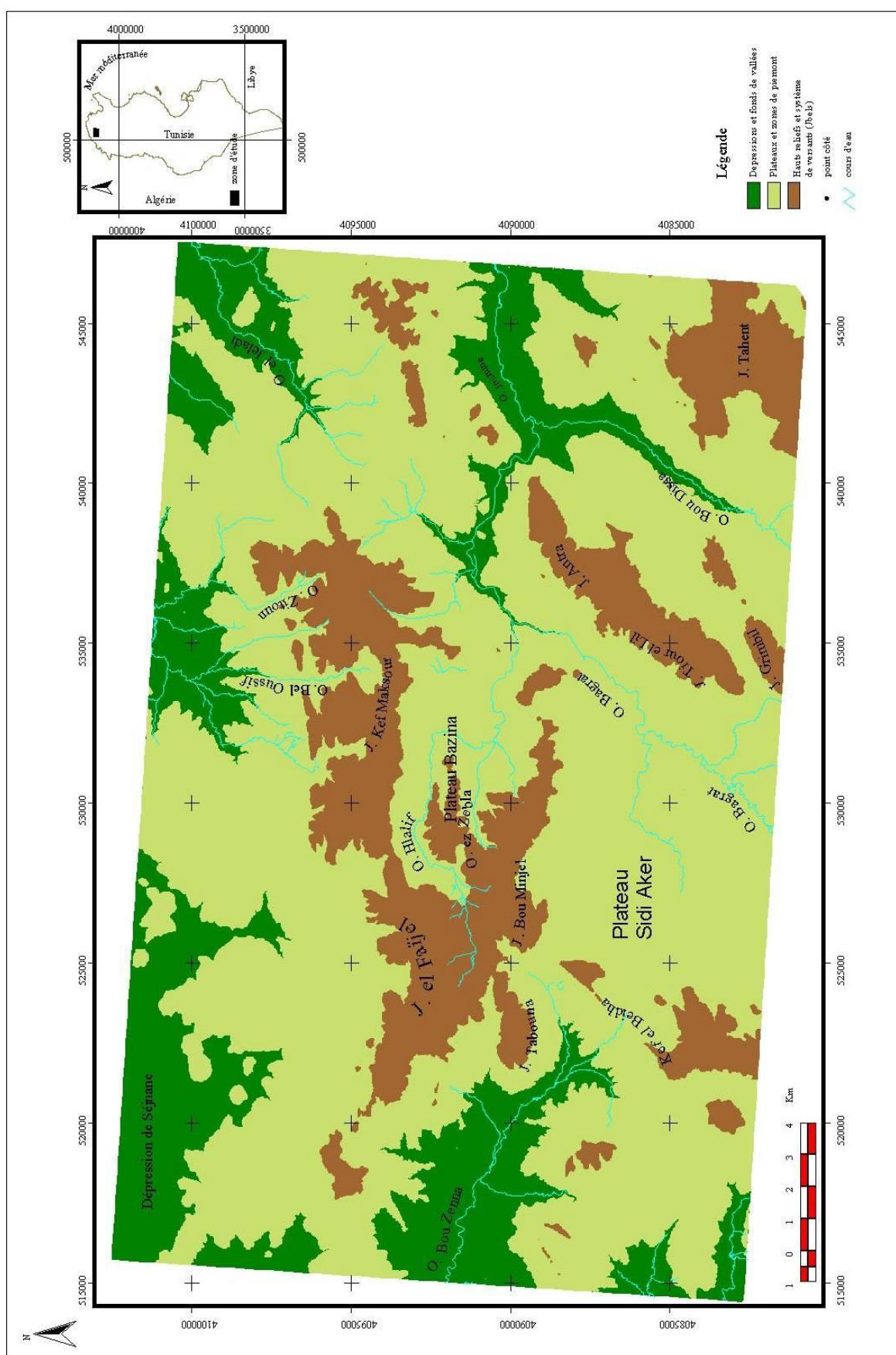
A partir du MNT des cartes dérivées, prenant en compte les données du relief, ont été réalisées : carte des classes d'altitude, d'ensoleillement et de pentes. Ces cartes peuvent être visualisées en plan, donc en 2 D ou "drapées" sur le modèle numérique de représentation du relief. Nous leur avons associé, comme pour n'importe quelle autre couverture, d'autres couches d'information (réseau hydrographique, la tectonique et la lithologie)

- la carte hypsométrique

La carte de classification des altitudes (Fig. 2a) est obtenue grâce à la méthode de classification par les écarts-type, supervisée par notre connaissance de terrain. Cette carte a été réalisée après visualisation de l'histogramme des fréquences relatives (Fig. 2b), qui présente un profil en cloche légèrement déporté vers les hautes altitudes. L'analyse de cette cartographie a fait ressortir les grands ensembles orographiques de la région.

- les cartes d'ensoleillement

L'illumination consiste à simuler un "ensoleillement artificiel" de la surface topographique érigée en 3D. Pratiquement on détermine les deux paramètres pour une source de lumière virtuelle à savoir l'azimut exprimé en degré (positif) de 0° à 360° à partir du Nord et de l'altitude de cette même source de 0° à 90° (0° degré correspond à l'horizon, 90° la position zénithale.





Parmi les autres discontinuités en relation avec les hauts reliefs, on cite le Jbel El Faidjel qui est dans la prolongation Ouest du Kef Maksour. Il diffère de celui-ci par sa lithologie. En effet, il est

taillé dans des dolomies du Trias et sa position, en bordure du complexe triasique, plaide pour l'hypothèse d'un diapirisme. Au centre-ouest de la carte, nous observons, surtout sur la figure 3, la forme subcirculaire du Jbel Tabouna avec un aspect dissymétrique plus connu sous le nom de demi-dôme et bien cerné du côté Nord par une faille. Le tracé du réseau hydrographique est, quant à lui, de type radial. A droite de cette montagne, s'allonge le Jbel Bou Menjel qui, comme son nom l'indique (faucille), dessine un arc d'orientation ouest-est dont les couches sont redressées à plus de 70° formant une barre. Ce Jbel, est encore une fois témoin du diapirisme responsable de cette topographie de plateau de Bazina qui joue le rôle d'un château d'eau dans la région.

La figure 4, met en relief d'autres discontinuités d'orientation SO-NE. Les plus manifestes sont celles qui affectent la zone des écaïles dans le quart SE de la carte. Celles-ci correspondent à une famille de failles signalées sur la carte géologique. On observe aussi le caractère disloqué de la « chaîne » de Grimbil-Antra. Cette « chaîne » est composée de deux anticlinaux parallèles mais qui présentent dans le détail beaucoup de nuances. En effet l'observation du MNT ainsi que de cartes dérivées montre que, à la différence de la carte géologique, ces deux ensembles sont loin d'être homogènes. Les deux jbel renferment des espaces de faible pente

qui correspondent aux zones érodées par le ruissellement qui a mis à profit soit une faiblesse lithologique ou exploité une fracturation. L'exemple type est celui de la cluse taillée par oued Joumine.

B – Les discontinuités de talwegs

La figure 3, montre des cours d'eau encaissés de direction E-O. Parmi ces oueds, on voit se profiler presque parallèlement oued El-Hlalif et oued Ez-Zebila au centre du plateau de Bazina. Nous remarquons aussi un changement spectaculaire du sens de l'écoulement de l'oued Bagrat, au niveau de sa confluence avec l'oued Es Sakka, au cœur de plateau de Sidi Aker. Cette déviation peut être d'origine tectonique, d'ailleurs un peu plus au sud de cette confluence une faille est signalée de direction parallèle au changement du tracé du cours d'eau.

La figure 4, nous permet de distinguer les plus importantes vallées de direction SO-NE. On cite l'exemple de l'oued Bagrat et celui de l'oued Bou Dissa qui alimentent l'oued Joumine. Cette orientation semble marquer les grands linéaments par l'importance des discontinuités mise en évidence par la carte d'ensoleillement. La superposition de la couche du réseau hydrographique et celle du réseau de failles sur le MNT fait ressortir encore plus les différentes discontinuités du tracé des thalwegs et des lignes de crête (Fig. 5 et 6).

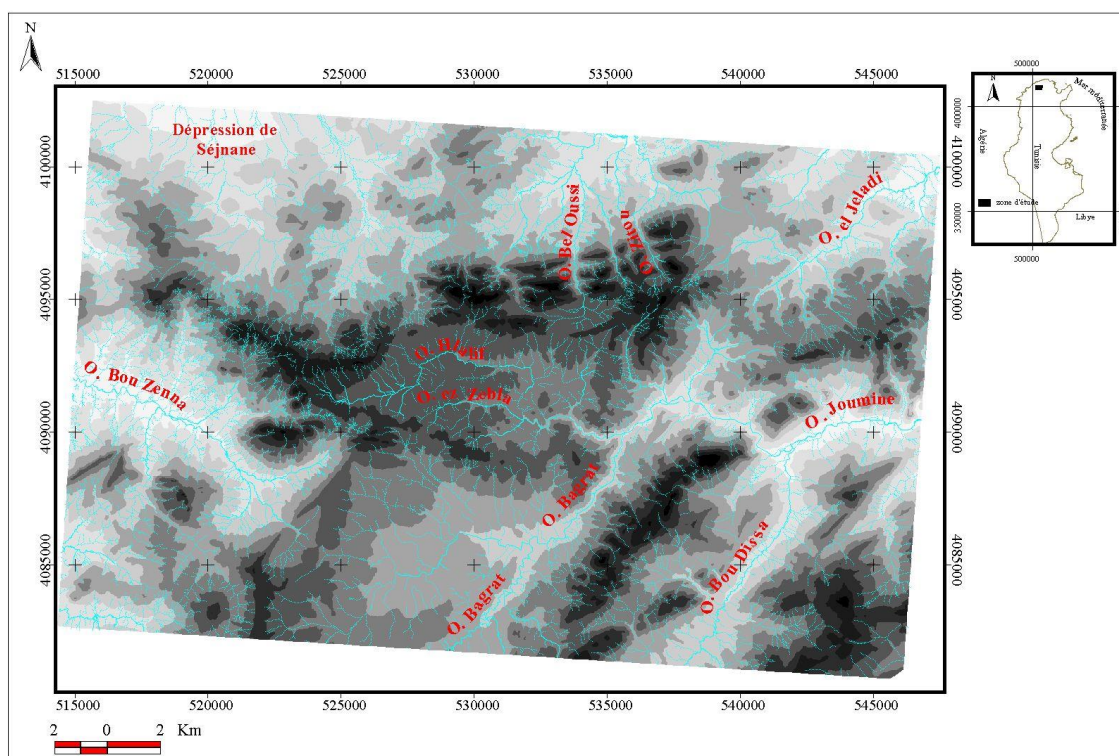


Fig. 5. Carte de superposition du réseau hydrographique avec le MNT

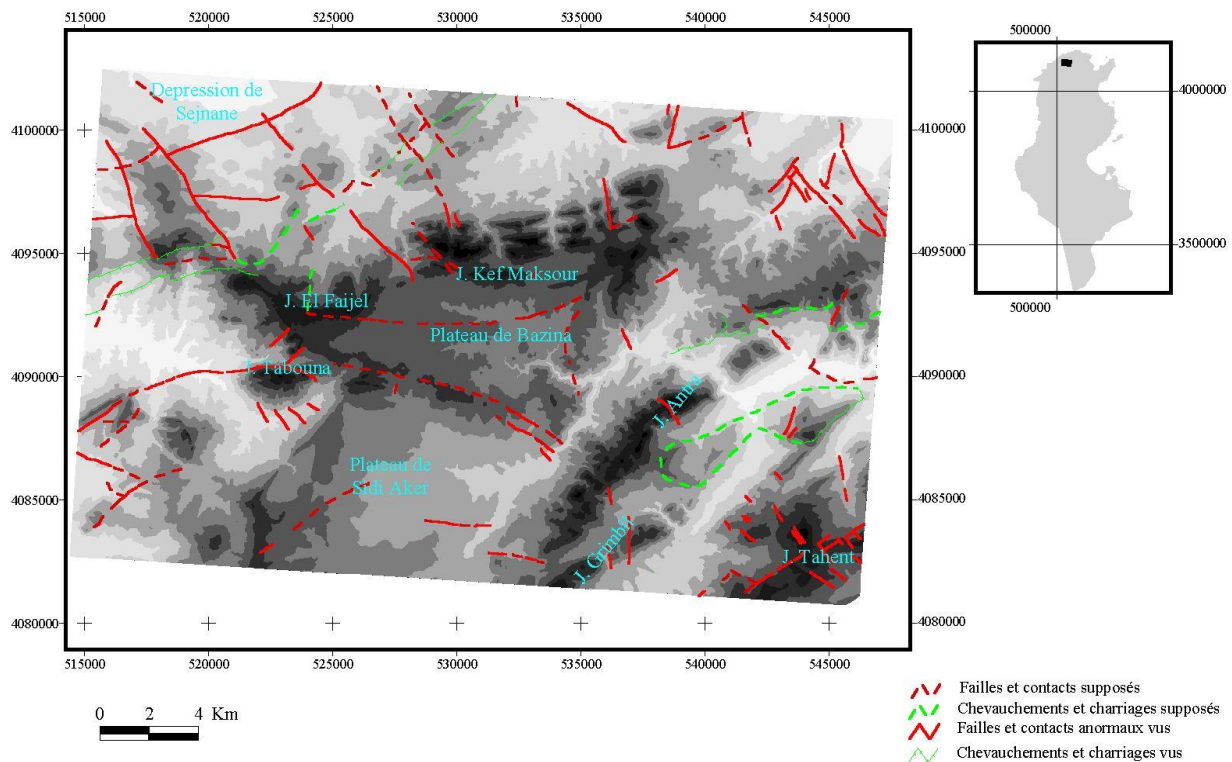


Fig. 6. Carte de superposition du réseau de failles et contacts anormaux avec le MNT

Par ailleurs, la comparaison de la rosace des directions du réseau hydrographique global (tout type d'écoulement confondu permanent ou non) et celle des directions de failles (vues ou supposées), montre une nette similitude (Fig. 7). En effet, quelque soit le critère retenu, soit la répartition en terme de fréquence de segments soit leurs longueurs cumulées, on remarque la prédominance de la direction N125 à N155. Cette même tendance se manifeste plus clairement si on se limite aux cours d'eau non permanents. Cependant, les cours d'eau permanents qui sont plus marqués dans le paysage montrent une direction globale N45 à N75. Cette direction est plutôt conforme à l'orientation des grands plissements dans la région. De plus, les deux oueds Bagrat et Bou Dissa coulent parallèlement aux anticlinaux des Jbel Antra et Grimbil.

En plus, des données de la tectonique, les données lithologiques font, elles aussi, sentir leur

empreinte dans la superposition du réseau hydrographique à la carte des affleurements géologiques (Fig. 8). Nous remarquons deux zones, l'une ayant une forte densité et l'autre ayant une faible densité hydrographique.

La zone de forte densité du réseau hydrographique observée dans le domaine des marnes paléocènes au nord de l'oued Bouzenna, traduit un paysage du type bad-lands. Ceci explique le caractère imperméable du substratum et sa prédisposition à l'érosion hydrique favorisée par une forte pente.

Ainsi la zone de faible densité du réseau hydrographique observée sur le plateau calcaire de Sidi Aker, s'explique par la perméabilité du substratum et la faiblesse de la pente qui ont favorisé des modelés karstiques.

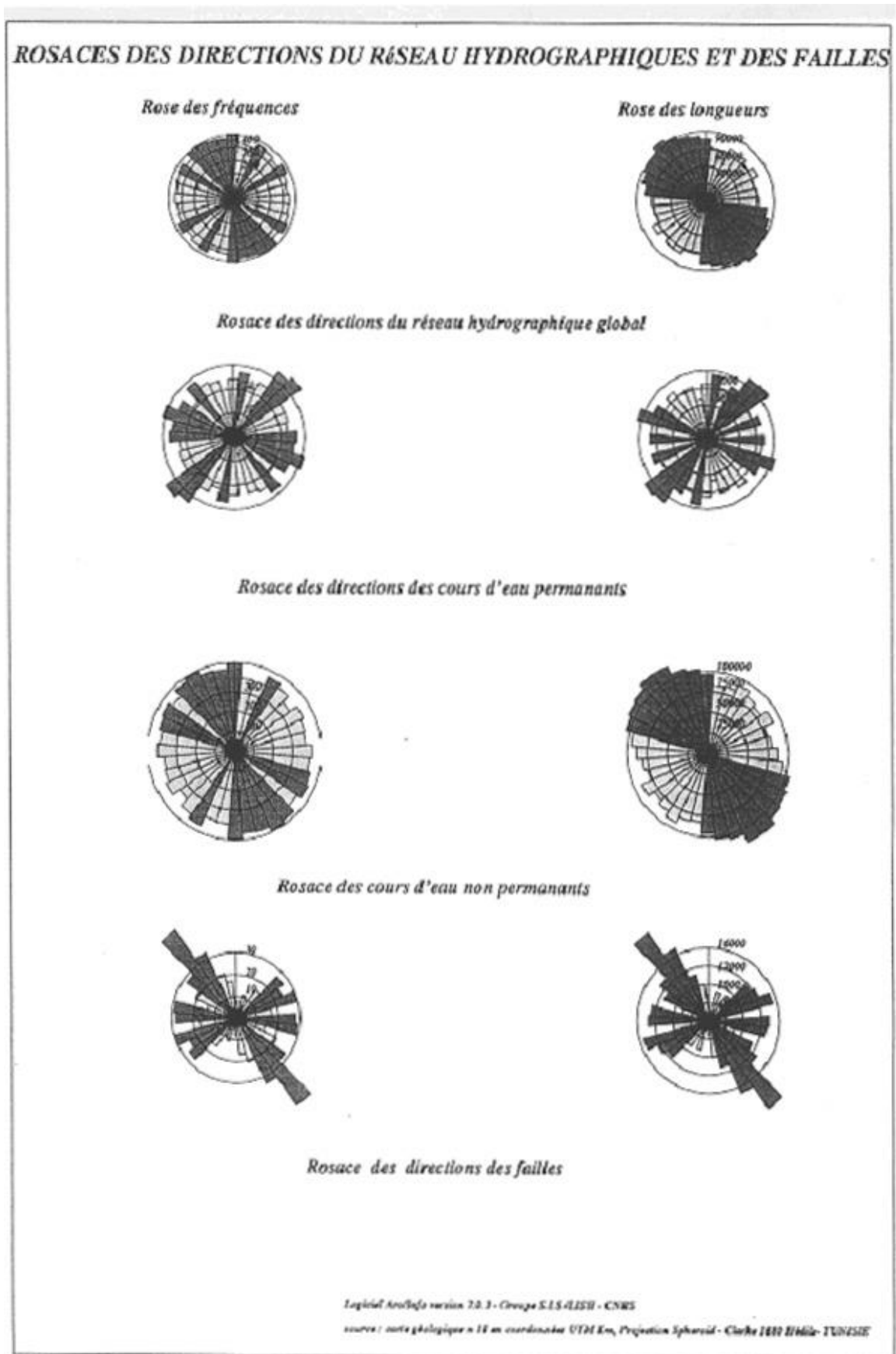


Fig. 7. Rosace des directions du réseau hydrographique global (tout type d'écoulement confondu permanent ou non) et celle des directions de failles (vues ou supposées)

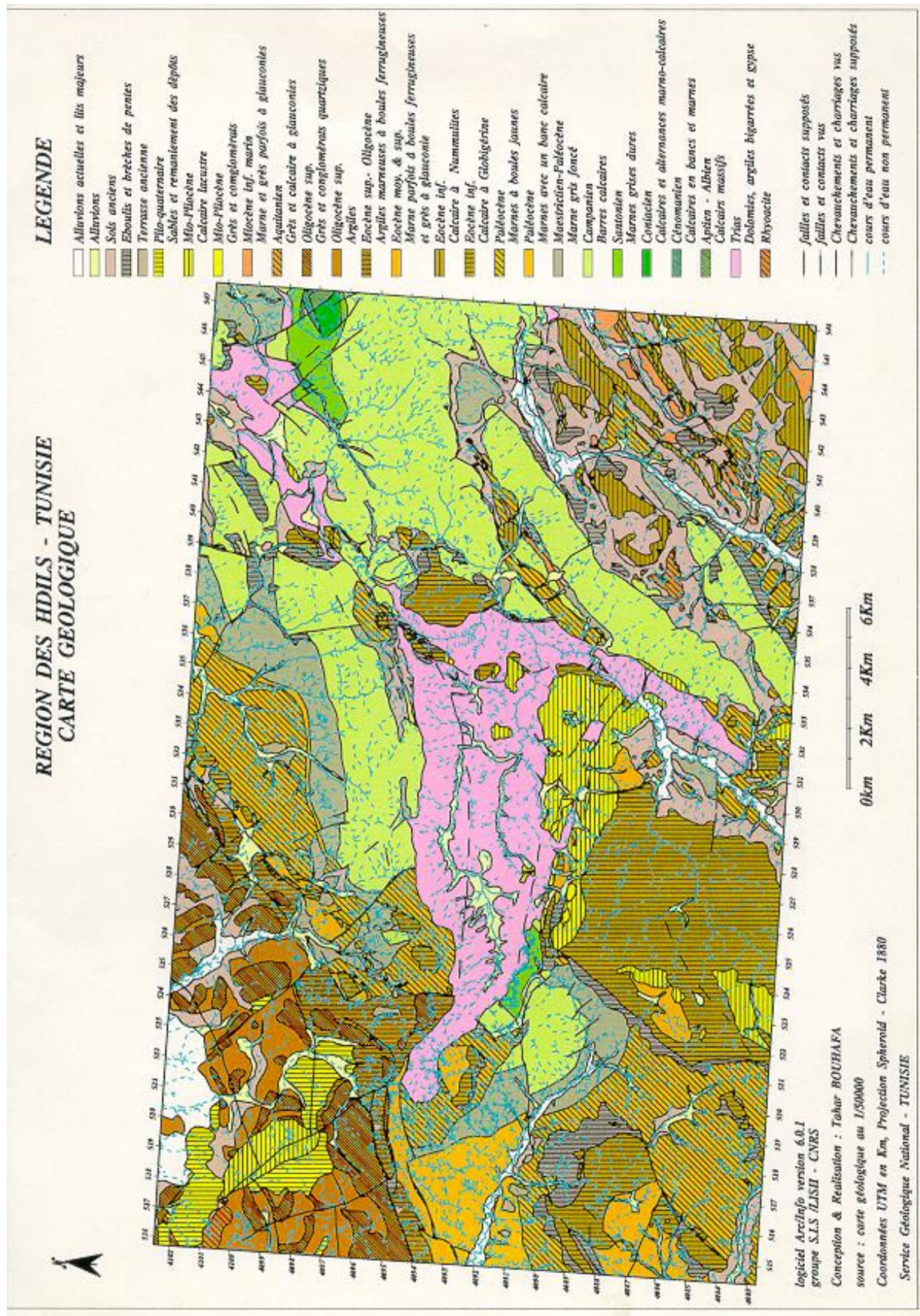


Fig. 8. La superposition du réseau hydrographique à la carte des affleurements géologiques

Conclusion

Les résultats de ces premières investigations soulignent que les éléments linéaires de la géomorphologie structurale peuvent être dégagés en utilisant une base de données géoréférencées et multicouches. Mais dans l'état actuel de nos recherches, nous constatons qu'un manque d'information peut handicaper le traitement, tel est le cas de la recherche de corrélation entre le tracé du réseau hydrographique et celui du réseau de faille. Cette constatation conduit à signaler que l'utilisation du SIG ne sert pas seulement à restituer de l'information à partir des données existantes mais aussi, c'est un moyen de prospection puisqu'il nous interpelle à partir de la détection de situation

d'anomalie (correspondance de la direction dominante des failles avec les cours d'eau non permanent).

La base de données Hédil-SIG, bien que l'information soit incomplète, garde l'intérêt d'associer les différentes couches d'information sous un format numérique et dans un système géoréférencé présentant l'avantage d'une prédisposition à un traitement automatique de l'information spatiale. En outre, dans l'attente de ce supplément d'information, il semble d'ores et déjà opportun de généraliser l'utilisation de l'approche SIG pour essayer de cartographier et donc d'interpréter les autres éléments de la morphologie structurale d'expression planaire.

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***Unité de Recherche Géomatiques des Géosystèmes 02/UR/10-01-Université de la Mannouba**

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Landscape Metrics for Assessment of Mountain Landscape using GIS Applications*

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Key words: landscape assessment, landscape metrics, Prahova Valley

Abstract: Using landscape metrics, we analyze the dynamics of landscape structure and landscape functionality of an area situated at the contact between the mountainous and sub-mountainous regions of the Prahova Valley.

1. Introduction

Sustainable landscape management represents a priority objective of environmental policy. Hence, the implementation of the European Landscape Convention aims at the identification, analysis and assessment of landscape. A better and more accurate assessment of a given area and of its dynamics can be obtained using indicators (e.g. landscape metrics), which quantify landscape structure or pattern. Landscape metrics measure two fundamental aspects of landscape structure: *composition* and *configuration* (Botequilha Leitão et al., 2006). They can be computed at patch level, for a class of patches or for the entire landscape mosaic. The composition metrics describe the variety and abundance of patches, without any reference to their spatial distribution. They measure the number of different land cover types (patch richness), the number of patches, the surface covered by each type (class area proportion) or several diversity indices (e.g. Shannon's and Simpson's diversity indices). Although not spatially explicit, they still have important spatial effects (Gustafson, 1998). On the other hand, the configuration metrics refer to the spatial arrangement and orientation of the patches in the landscape mosaic. These metrics give information about the patch shape, the distance between classes of same type (nearest neighbor distance) or the degree of contrast between patches along patch edges (edge contrast). The aim of our study is to apply this method for landscape assessment of several mountainous and sub-mountainous regions in Romania. For the pilot study, we have chosen an area situated in the region of the Prahova Valley and we computed several landscape indicators corresponding to the years 1989 and 2007.

The study area is located at the contact between the mountainous and sub-mountainous regions of the Prahova Valley (Comarnic-Posada) and has a surface of approx. 6 km². We have chosen this region, due to its landscape complexity and pronounced dynamics. The increasing of the real estate investments, the development of the touristic resorts, the Bucuresti-Brasov highway project prove that the Prahova Valley is a region which is permanently modifying its landscape structure and characteristics. Understanding the landscape dynamics for this area could be relevant in order to prioritize the anthropic interventions in this region. In this way, one could prevent the landscape degradation and losing natural, cultural, national or European heritage values.

3. Methodology

3.1 Data collection, data optimization, map production and data analysis

We used maps at a scale of 1:50,000, which were acquired for the year 1989 (Fig. 1). These maps were scanned at 600 dpi and then geo-referenced using Image Analysis – Arc View GIS software (version 3.1) in Stereo70 projection. We also used aerial photographs at a scale of 1:5,000, which were acquired in the year 2007 (Fig. 2). These photographs were registered to a rectified image. In order to obtain a more accurate land-cover-type identification, the photo-interpretation process involved stereoscopic viewing. In the map production system we used geo-database, which

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included feature classes with many elements. Land-cover-type boundaries were digitized using Arc View GIS software (version 3.1). The software was equipped with a symbol library. This library was necessary to represent all land-cover-types. The vector files resulted by the digitization process were

transformed into a raster format (grid image). These grids were analyzed with the FRAGSTATS software (version 3.3) developed by McGarigal et al. (1995), obtaining several landscape metrics for the study area.



Fig. 1. Topographic map at a scale of 1:50,000 (edition 1989)

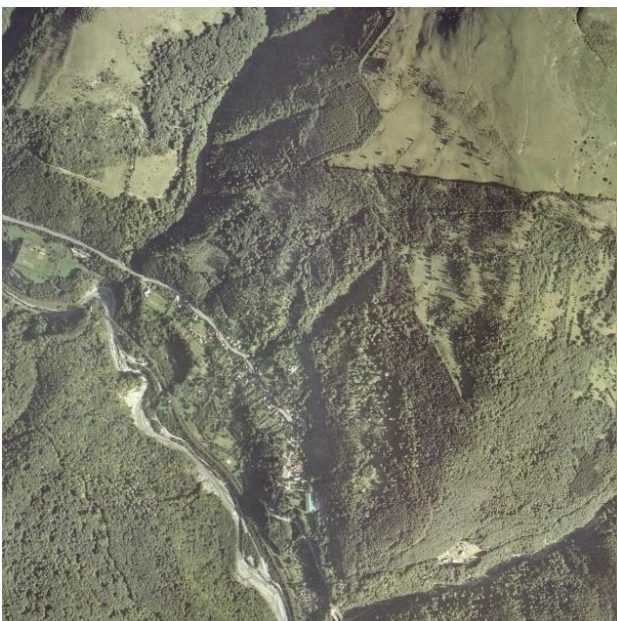


Fig. 2. Aerial photographs at a scale of 1:5,000 (edition 2007)

3.2. Landscape metrics

We selected for our study the following landscape metrics: Patch Richness (PR), Patch Number (PN) and Shape ($SHAPE$). Let us briefly recall the way they are defined and their meaning. The first metric chosen for our study is a composition metric, namely *Patch Richness* (PR). It represents the number of different land cover types (classes) and it is given by the formula:

$$PR = m,$$

where m = number of classes present in the landscape (McGarigal, Marks, 1995). The second metric analyzed in our research is *Patch Number* (PN). It measures the landscape configuration and it is given by the total number of patches:

$$PN = \sum_{i=1}^{PR} P_i,$$

where P_i = number of patches of type i . The third metric is *Shape* ($SHAPE$), which is again a configuration metric. At the patch level, it is a measure of the geometric complexity of the patch. The formula which defines this metric is:

$$SHAPE = \frac{p_{ij}}{f(p_{ij})}$$

where p_{ij} = perimeter (m) of patch j of type i and the function f is given by

$$f(p_{ij}) = \begin{cases} 4n, & \text{when } a_{ij} = n^2 \\ 4n + 2, & \text{when } n^2 < a_{ij} \leq n^2 + n \\ 4n + 4, & \text{when } a_{ij} > n^2 + n \end{cases}$$

Here n denotes the largest square number smaller or equal to the area a_{ij} of the patch. For the whole landscape, one can compute the mean value, given by the relation

$$SHAPE_{MN} = \frac{\sum_{j=1}^{PN} SHAPE_j}{PN}$$

where $SHAPE_j$ represents the value of the landscape metric $SHAPE$ corresponding to the patch j .

4. Results

For the first metric, Patch Richness, we obtained $PR_{1989} = 4$, respectively $PR_{2007} = 5$. More precisely, the landscape mosaic contained the following land-cover-types in the year 1989: forest, pastures, urban and open areas. These classes were also present in 2007 and a new class appeared: cleared forest, which represents 0.46 % of the total area. For the second selected metric, Patch Number, the results are $PN_{1989} = 27$, respectively $PN_{2007} = 46$. The metric $SHAPE$ varies between 1.000 and 3.389 in 1989 and between 1.000 and 4.166 in 2007. For the whole landscape, the mean values are $SHAPE_{MN_{1989}} = 1.527$, respectively $SHAPE_{MN_{2007}} = 1.712$ (Fig. 3 and 4).

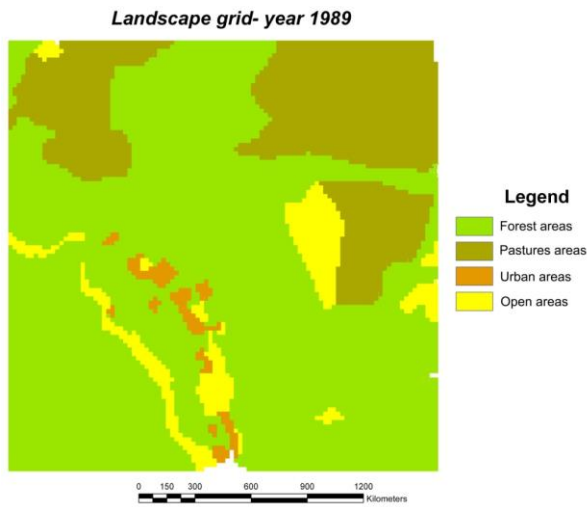


Fig. 3. Landscape grid map at a scale of 1:50,000 (edition 1989)

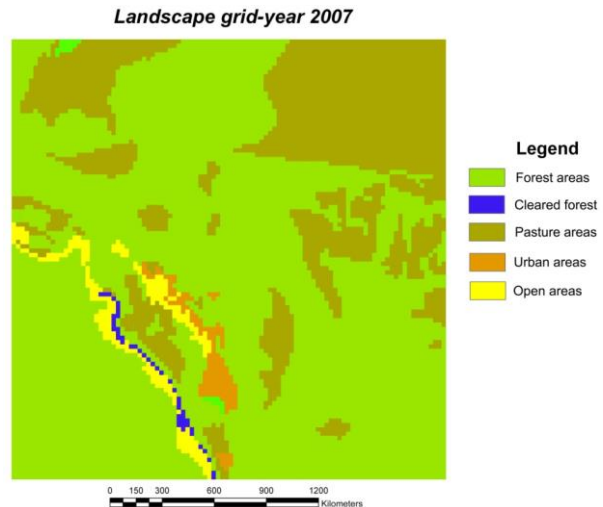


Fig. 4. Landscape grid map at a scale of 1:5,000 (edition 2007)

5. Conclusion

The use of landscape metrics allowed us to point out more accurately the changes occurred in the landscape structure and in its functionality. We first observe the diversification of the landscape cover types and the growth of the fragmentation degree. An important change of the landscape structure is the appearance of cleared forest, conjugated only with local reforestation. This fact shows that one should adopt efficient deforestation control policies.

Concerning the landscape functionality dynamics, we notice that several forest and pasture areas changed their destination, becoming urban areas.

The research will continue by investigating other areas in the same region. We conjecture that the increasing of the fragmentation degree is quasi-ubiquitous in the Prahova Valley. It would be interesting to study the effects of this phenomenon on different wildlife species, on ecological processes or on the quality of people's life.

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Apport de l'imagerie radar SAR (Images d'Amplitude) pour l'analyse des aléas et du changement dans la région de Buzău (Roumanie)*

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Mots clefs : Radar, Image Amplitude, Composition colorée, SAR, Analyse diachronique.

Résumé. La région de Buzău, située sur la région subcarpatique en Roumanie est le siège de deux grands types d'aléas naturels : des aléas géologiques-geomorphologiques (glissements de terrains ; séismes dans sa partie Nord-Nord Est ; des volcans de boue sont également présents sur la zone) et des aléas climatiques ou hydrologiques (phénomènes d'inondation accentuant souvent l'érosion des sols) (Grecu et al., 2007).

On peut alors s'interroger sur le moyen le plus adapté pour établir un suivi permanent de ces risques, en établir une cartographie aussi fidèle que possible et estimer, éventuellement, les dégâts et les modifications du paysage qui en résultent. L'imagerie satellitale, avec ses diverses résolutions, apparaît comme l'un des moyens les plus appropriés, sinon l'incontournable, pour résoudre le problème ainsi posé. A cet effet, Abdellaoui et al. (2005; 2007) ont essayé d'analyser l'apport des images optiques pour le suivi du risque naturel en Roumanie.

De la télédétection active ou passive, l'un des avantages des images radars est la possibilité d'enregistrement par tout temps indépendamment des conditions météorologiques (jour, nuit, nuages). Cette capacité leur donne un avantage de leur exploitation, pour le suivi des phénomènes en fonction du temps. La radiométrie enregistrée est fonction, entre autres paramètres, de l'humidité et de la rugosité de la surface ; ainsi les surfaces rugueuses et humides sont représentées par une forte radiométrie ; au contraire les surfaces lisses et arides apparaissent avec des valeurs faibles de radiométrie.

Une étude multi date peut être réalisée à partir d'images composites associant des images d'une même scène à deux dates différentes. La composition colorée diachronique met ainsi en évidence les différences entre deux dates ; mais ces différences ne sont pas toujours causées par une variation des propriétés de la surface imagée. D'autres processus peuvent être à l'origine de ces variations ; nous citerons le recalage des images ou les conditions géométriques d'acquisition.

Pour le présent travail, nous avons associé plusieurs couples d'images séparées par des intervalles de temps différents pendant lesquels nous avons essayé de détecter et d'interpréter les changements. Nous avons pour cela utilisé des images des satellites ERS et ENVISAT acquises par le laboratoire de géomorphologie de l'université de Liège dans le cadre d'une recherche partagée entre les universités Paris12, Liège et Bucarest, financée par l'AUF ; nous avons produits des images composites associant des scènes séparées par des intervalles de 1 à 1646 jours.

Les images utilisées couvrent une superficie de 100km x 100km.

L'analyse diachronique a permis de mettre en évidence les variations de surface, notamment au niveau des volcans de boue, des lacs (variation du volume d'eau présent), de la plaine (variation de la couverture végétale).

1. Introduction

L'image radar est une image complexe. La mesure associée à chaque pixel est constituée de deux parties : une partie réelle et une partie imaginaire. A partir de ces deux paramètres les deux informations d'amplitude et de phase sont extraites. La phase est exploitable indirectement par la technique d'interférométrie à l'aide d'une autre phase de la même scène. L'amplitude caractérise la réflectivité du pixel pour l'onde hyper fréquence. Elle est

exploitable directement et comparable à l'image optique. Elle est en relation directe avec l'état de surface ; ainsi les surfaces rugueuses et humides sont représentées par une forte radiométrie ; au contraire les surfaces lisses et arides apparaissent avec des valeurs faibles de radiométrie.

Les images radars d'amplitude ont été utilisées dans plusieurs domaines, comme la cartographie du dommage urbain ou aux catastrophes naturelles ou industrielles, les changements des limites côtières

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(Bâ et al., 2007), le suivi de la déforestation, la surveillance de la fonte des glaces, la lithologie dans l'environnement polaire (Budkewitsch et al., 1996a, b), la cartographie géologique (Singhroy et Saint-Jean, 1999), la cartographie structurale (Wade et al., 2001), la classification des cultures (Bruniquel et Lopès, 1994 ; Lopès et Séry, 1997), la reconnaissance des formes, l'aménagement du territoire (Rudant et al., 1996), la détection et la caractérisation des nappes d'hydrocarbure dans les mers et les océans (Mercier et al., 2004). D. Troufleau et al. (1994), ont associé l'image amplitude au MNT pour montrer l'efficacité de ce type d'image pour déterminer l'humidité de surface du sol en zones arides et semi arides. Dans l'exploration géologique en zones arides, N. Baghdadi et al. (2005) ont démontré que l'imagerie radar SAR, en bande « L » ou « C », montre un plus fort potentiel de cartographie que les images optiques ASTER, surtout pour les structures linéaires comme les dykes ainsi que les structures circulaires. P. Laborde et M. Deveau (1996) ont étudié les capacités et principales limites de l'imagerie SAR/ERS-1 pour la détection de phénomènes méso échelle en océanographie ; ils ont montré que seuls les courants supérieurs à 10m/s sont détectables par vent de vitesse supérieure ou égale à 10 m/s.

Les images radars d'amplitudes, constituent un outil très efficace pour le suivi dans le temps des phénomènes hydrologiques (les crues, les inondations, etc.). Elles permettent d'évaluer l'impact des inondations sur le territoire. Cependant une image satellitaire seule ne permet pas de comprendre pourquoi certaines zones sont inondées et d'autres pas ; elle ne permet pas de faire de prédictions quant aux terrains qui seront envahies par les crues et d'identifier des zones qui peuvent servir pour une évacuation rapide. Il faut l'associer à d'autres images et un MNT. P. A. Brivio et al. (2002) ont pu cartographier l'extension de l'inondation au moment de l'acquisition, à l'aide de deux images radars SAR du satellite ERS-1 acquises un mois avant l'inondation et trois jours après l'inondation en novembre 1994 dans la région de Piémont en Italie. A. McMillan et al. (2006) ont utilisé une composition colorée pour l'identification des caractéristiques optimales pour la surveillance de l'inondation dans des zones urbaines. Dans le domaine de la géomorphologie, les images radars permettent de détecter des zones humides susceptibles de connaître des glissements ; cependant, on ne peut pas prévoir avec exactitude le

sens du glissement ni le volume des blocs déplacés ; le sens du glissement pourra être déduit à partir d'une carte de pentes, elle-même produit dérivé des MNT. Pour l'estimation du volume des blocs déplacés, les cartes géologiques et géomorphologiques sont nécessaires.

L'objectif de ce travail est de montrer l'apport de l'imagerie radar pour l'analyse des changements du paysage dans la région subcarpatique de Buzau (Roumanie), en utilisant l'amplitude du signal radar rétrodiffusé entre différentes dates d'acquisitions. Nous avons associé plusieurs couples d'images séparées par des intervalles de temps différents pendant lesquels nous avons essayé de détecter et d'interpréter les changements. Nous avons pour cela utilisé des images des satellites ERS et ENVISAT acquises par le laboratoire de géomorphologie de l'université de Liège dans le cadre d'une recherche partagée entre les universités Paris 12, Liège et Bucarest, financée par l'AUF ; nous avons produits des images composites associant des scènes séparées par des intervalles d'1 à 1646 jours.

2. Méthodologie

2.1. L'analyse multi date

Pratiquement, la composition colorée diachronique met en évidence les différences entre deux dates ; mais ces différences, pour l'image radar, ne sont pas toujours causées par la seule variation des propriétés de la surface imagée. D'autres processus peuvent être à l'origine de ces variations ; nous citerons le recalage des images ou les conditions géométriques d'acquisition.

On peut alors modéliser le changement produit entre deux acquisitions par une fonction de trois variables (trois sources de natures différentes) : la scène imagée, la géométrie et le bruit ; on obtient ainsi :

$$\zeta_{chang}(p) = f \left[\zeta_B(p) \cdot \zeta_G(p) \cdot \zeta_S(p) \right]$$

$\zeta_B(p)$: traduisant les erreurs de calage, de traitement ...

$\zeta_G(p)$: pour les décorrélation causées par la différence des conditions géométriques de prise vue (différents angles d'incidence, ligne de base trop grande, différentes polarisations)

$\zeta_S(p)$: Du aux caractéristiques physiques ou diélectriques du pixel observé (c'est le changement recherché).

Cette décomposition du changement observé (par séparation des causes probables) est utile pour l'interprétation des images composites d'amplitudes et pour estimer les changements relatifs à l'évolution des caractéristiques propres des zones étudiées.

Cependant, une analyse multi temporelle à partir d'images amplitudes n'est possible qu'à certaines conditions :

- Dans une analyse multi date, les données doivent être en provenance du même centre de traitement et d'archivage (PAF) ; en effet, chaque PAF a des algorithmes propres de traitement de signal.
- La période séparant deux acquisitions doit être cohérente avec la période moyenne de changement pour le phénomène observé.
- L'angle d'incidence doit être identique pour les deux acquisitions, car il est difficile de comparer ou de superposer des images acquises sous des incidences différentes. La radiométrie, dans l'imagerie radar, est en effet très sensible à l'incidence de prise de vue. Pour les applications multi temporelles, la comparaison avant et après un événement, nécessite de travailler à incidence unique. (Adragna, 2001). Cela exclut les images acquises en orbite de sens opposés (ascendant et descendant) ; dans ce cas en effet, il est très difficile de corréliser les images ; les objets (bâtiments, arbres, montagnes, etc.) sont vus par leurs faces opposées et apparaissent alors de façons très différentes.
- Enfin, les deux images doivent être prises avec un capteur utilisant la même longueur d'onde. En effet la rugosité relative d'une surface par rapport à l'onde radar dépend de la longueur d'onde et de l'angle d'incidence. Le travail de P. Laborde et M. Deveaux, (1996), sur l'uniformisation de l'image à cause de la variation de l'angle d'incidence de l'onde émise entre 19° et 26°, montre qu'il y a une diminution de l'amplitude du signal radar entre les points allant de la portée proximale (points plus proches de l'antenne) vers la portée distale (points plus éloigné de l'antenne). Cette variation se traduit par une baisse d'intensité des niveaux de gris de l'image selon l'axe radial. Si on applique le critère de Rayleigh,

$$\Delta h > \frac{\lambda}{8 \cos \theta}$$

pour deux angles d'incidence (19° et 26°) dans le cas du satellite ERS et ENVISAT, pour une longueur d'onde $\lambda = 0,0565$ m, on obtient une valeur Δh comprise entre 7,46901381 mm pour une incidence de 19° et 7,85686985 mm pour une incidence de 26° soit environ 0,40 mm de différence ; pour 23° on obtient la valeur $\Delta h = 7,67175761$ mm. Ceci montre qu'une différence de quelques degrés affecte la rugosité en fraction de millimètre, l'influence est donc minime.

2.2. Les données utilisées et la zone d'étude

2.2.1. La zone d'étude

La zone d'étude (Fig. 1) est un espace de 100km x 100km (couverture des images radar) englobant le comté de Buzău et des parties d'autres comtés (Vrancea, Braila, Prahova et Covasna). Elle subit deux types de risques naturels particuliers: les risques climatiques (phénomènes d'inondations; citons en particulier celles du 7 au 31 mai 2005 causant des dégâts matériels estimés à 150 millions d'euros) et les risques géologiques (tremblements de terre, glissements de terrain et coulées de boue). Les précipitations de fortes intensités, associées à des périodes de sécheresse, accentuent l'érosion des sols et favorisent le phénomène de glissement de terrains.



Fig. 1. Localisation de la zone de travail

2.2.2. Les données images

Les données SLC utilisées, ont été traitées par compression d'impulsion en direction radiale en distance et par la synthèse d'ouverture RSO (SAR) en direction azimutale à partir du même centre de traitement et d'archivage (UK-PAF).

Pour cette étude, nous avons utilisé 8 images :

– 4 images mono vues complexes : une image du satellite ERS-1 et trois images du satellite ERS-2, aux dates : 28 et 29 mai 1995, 17 Juin 1996 et 29 novembre 1999;

– et 4 images ENVISAT : deux en 2004 (octobre et décembre) et deux en 2005 (janvier et avril).

Ces images ont respectivement 26460 lignes et 4900 colonnes pour ERS et 26890 lignes et 5178 colonnes pour ENVISAT. Elles couvrent une surface de (100km x 100km) avec une résolution de 4m (en azimuth) et 20m (en distance). Le tableau (1) synthétise les caractéristiques de ces images.

Tableau 1 : Images Amplitude			
Dates	Orbites	DATUM	Centre image (Lat. / Long)
ERS			
28-05-1995	20221	GEM6	(45,377/ 26,802)
29-05-1995	548	GEM6	(45,387/ 26,807)
17-06-199	6059	GEM6	(45,386/ 26,806)
29-11-1999	24095	GEM6	(45,385/ 26,832)
ENVISAT			
18-10-2004	13774	WGS84	(45,404/26,895)
27-12-2004	14776	WGS84	(45,419/ 26,904)
31-01-2005	15277	WGS84	(45,434/26,911)
28-05-1995	16279	WGS84	(45,460/ 26,916)

2.3. Obtention des images amplitudes

2.3.1. Extraction des images amplitude

A l'aide des logiciels SAR, DIAPASON et DORIS, nous avons extrait les images amplitudes ramenées au format multivues par un rééchantillonnage spatial de 5 pixels en azimuth et 1 pixel en distance afin d'avoir une dimension carrée du pixel ainsi qu'une résolution maximale de 20m dans les deux directions (azimutale et distance). Ceci a aussi pour conséquence de diminuer le chatoiement qui caractérise les images amplitude radar SAR. Elle est réalisée par la sommation incohérente de plusieurs vues en intensité (Grandchamp et Cavassilas, 1997). Nous disposons ainsi de quatre images amplitudes radar SAR de taille (5292 x 4900) pour ERS et quatre images de taille (5378 x 5178) pour ENVISAT.

2.3.2. Calcul de coefficient de rétrodiffusion et recalage des images

Les images amplitudes ne sont pas corrigées et ne sont pas données avec leur coefficient de rétrodiffusion. Aucune opération de calibration n'est réalisée par le logiciel Diapason, orienté vers les études de phase et l'interférométrie DinSAR. Les images extraites gardent les informations d'amplitude propres au produit de base. De fait, aucun coefficient de rétrodiffusion ne peut être extrait directement. Nous devons pour cela utiliser le logiciel BEST¹, fourni par l'ESA. Cette étape consiste à convertir l'image d'amplitude extraite en une image d'amplitude avec comme information le coefficient de rétrodiffusion pour chaque pixel.

2.3.3. Filtrage des images amplitudes

La lisibilité des images radars d'amplitudes est limitée par la présence de chatoiement (ou speckle). Le *speckle* est un bruit granulaire dû au fait que chaque élément rétrodiffuseur élémentaire se comporte comme une source émettant une onde cohérente et qui peut inférer en tout point de l'espace. L'inférence peut être, soit une inférence constructive, où le résultat est une tache brillante constituée par des pixels clairs sur l'image, soit une inférence destructive, où le résultat dans ce cas est une tache sombre que forment les pixels foncés. Pour augmenter la lisibilité, il existe des filtres spécifiques comme les filtres basés sur des modèles statistiques (filtre Frost 1981, Lee 1986, filtre de Kuan 1987...). Pour ce travail, nous avons utilisé le filtre Frost pour toutes les images amplitude avec une fenêtre moyenne de (3x3).

2.3.4. Géoréférencement et calage

Cette étape a été très délicate. En effet, il n'a pas été facile d'identifier des pixels « semblables » dans les images à recalcr. Nous finalement du réaliser le calage des images amplitude l'une par rapport à l'autre à l'aide du logiciel ENVI en opérant séparément sur les images ERS d'une part et les images ENVI d'autre part.

Après géoréférencement nous avons encore constaté l'existence d'un décalage des images entre elles. Ce décalage est dû à la différence de géométries de prises de vue ainsi qu'aux différences d'orbites.

¹ <http://earth.esa.int/services/best/>.

3. Premiers résultats et interprétation

3.1. Lecture des images amplitude

Dans les images amplitudes, une surface lisse apparaît noire ou très sombre, à cause des variations moyennes des aspérités des pixels inférieures à 7.67mm suite à la relation qui existe entre la longueur d'onde utilisée et la rugosité. La réflexion spéculaire entraîne un faible retour d'énergie vers le capteur. La surface rugueuse, apparaît claire suite à la réflexion diffusée de l'onde incidente, caractérisée par la capacité de cette surface à renvoyer l'énergie vers le radar.

Les points les plus brillants dans l'image amplitude sont dus aux coins réflecteurs comme les bâtiments dans des zones urbaines ou les sommets des falaises en zone rurale ; les empilements de roches produisent de multiples réflexions surtout si l'angle droit est bien orienté vers le capteur. Dans notre zone d'étude, on distingue bien les villes grâce à leur capacité de rétrodiffuser le signal radar ; sur l'image de la figure 2, on voit nettement Buzău, Focșani, Raminicu-Sarrat, Mizil, Galati... ; les trois lacs apparaissent en noir signe d'une surface lisse.

La rétrodiffusion du signal par la scène imagée vers le capteur radar est influencée par l'inclinaison du sol et sa rugosité (définie par l'écart quadratique moyen des irrégularités d'une surface) ; plus le sol est rugueux plus il a une forte intensité et il apparaît alors clair dans l'image. Les caractéristiques diélectriques du sol influencent également le signal ; les objets qui contiennent beaucoup de métal réfléchissent fortement comme le cas ici à Buzău où 85% des maisons ont des toits en métal (zinc). Des anomalies peuvent également apparaître dans l'image à cause de l'effet de chatoiement.

Une différence d'angle d'incidence induit une différence du seuil de rugosité à détecter. Pour les satellites ERS et ENVISAT et avec bande « C », $\lambda=0,0565$ m :

- pour ERS : ($\theta_{ERS} = 23,3^\circ$) ; $\cos \theta = 0,91844$; une surface est rugueuse si elle vérifie : $\Delta h > 7.689\text{mm}$.

- pour ENVISAT : ($\theta_{ENVISAT} = 22,8^\circ$) ; $\cos \theta = 0,92186$; une surface est rugueuse si elle vérifie : $\Delta h > 7.661\text{mm}$.

L'humidité agit sur la transmission et l'absorption du signal. Plus la teneur en eau est élevée, plus l'intensité rétrodiffusée est forte. Ainsi, la constante diélectrique détermine la profondeur de pénétration de l'onde radar ; quand elle est élevée, l'onde pénètre moins dans le sol et l'intensité rétrodiffusée est forte. Une augmentation de taux d'humidité augmente la constante diélectrique et amplifie la brillance du sol. Un terrain sec, au contraire, aura un signal faible.

3.2. Le principe de la composition colorée

Nous avons réalisé les composites colorés par couple en affectant le vert à l'image la plus ancienne et le rouge à l'image la plus récente. Nous obtenons ainsi la couleur de chaque pixel de l'image résultante par la combinaison suivante :

$$c_{(p)} = a_{(p)} [V] + b_{(p)} [R],$$

où $a_{(p)}$ et $b_{(p)}$ représente respectivement des valeurs dépendant de l'intensité du signal du pixel pour la composante la plus ancienne et la plus récente.

Ainsi pour un pixel apparaissant en noir dans l'image 1 (intensité du signal égale à 0) et en blanc dans l'image 2 (intensité du signal égale à 1), nous avons :

$a_{(p)} = 0$ et $b_{(p)} = 1$ d'où $c_{(p)} = [R] \rightarrow$ le pixel apparaît « rouge »

Le tableau 2 suivant résume les cas extrêmes pouvant être obtenus

Tableau 2 : Les cas extrêmes des couleurs résultantes de la composition colorée			
a\b	0	$0 < \beta < 1$	1
0	Noir		Rouge
$0 < \alpha < 1$	Couleurs intermédiaires entre le vert et le rouge		
1	Vert		Jaune

3.3. Image amplitude

Sur les images amplitudes radar SAR, on distingue nettement trois parties bien différenciées : une partie plate d'altitude allant de 30 à 80m (agriculture et foyers urbains), une partie intermédiaire d'altitude moyenne allant entre 300 et 400m (zone de glaciaires)

avec une bonne réflexion suite à leur orientation vers la visée du radar et leurs très faibles pentes ($\alpha < 23^\circ$) et une partie de relief d'altitude allant de 850 à 1500m (partie des Subcarpathes et Carpathes Orientales de la région de Buzău). Cette partie comprend : la limite entre les Carpathes orientales et les Subcarpathes, les deux volcans de boue (Pâclele)

Mari et Mici) Sud et Nord, une faille, des zones de glissements de terrain, et la rivière de Buzau au Sud des deux volcans de boues.

Les principaux éléments du paysage sont bien identifiés (Fig. 2) ; en zone de plaine, les lacs sont bien visibles, notamment quand ils sont remplis d'eau en absence de vent « fort » (apparence en noir, caractéristique d'une surface lisse) ; les parcelles sont bien distinguées à cause de leur différence de

rugosité ; les villes apparaissent en blanc traduisant une forte rétrodiffusion du signal (les coins sont très réflecteurs ainsi que les toits des maisons à cause du matériau de couverture) ; en zone de relief, il y a également une forte rétrodiffusion du signal ; on notera également la déformation de l'apparence du paysage : en particulier apparence d'un « cordon » en dessous de la région des volcans de boue.

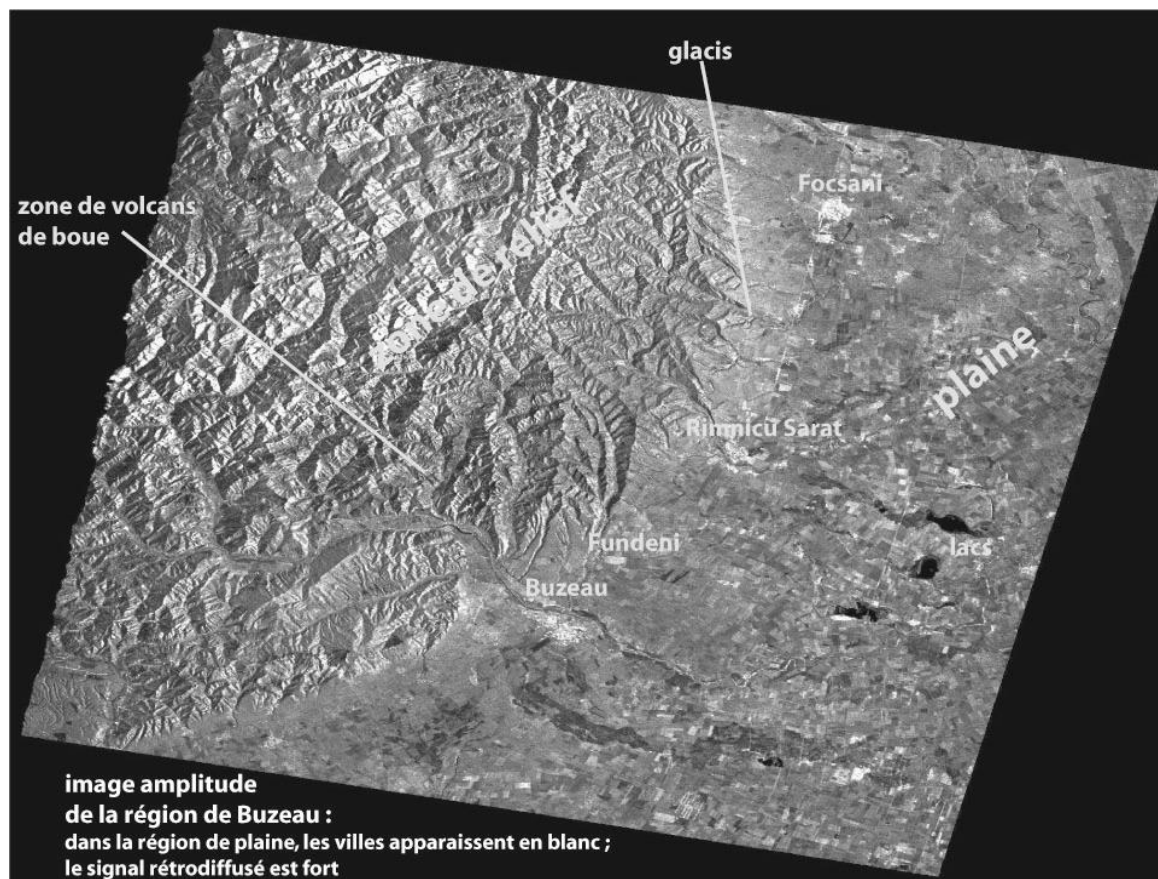


Fig. 2. Image amplitude de la zone d'étude

3.4. Validation

Sur les images amplitude réalisées, nous constatons une faible distorsion dans les deux dimensions du pixel, causée par la différence de résolutions dans les deux directions. On observe ainsi un léger allongement vers le Nord (direction azimutale). Cependant, les images sont relativement bien corrélées ; pour le confirmer, nous observons des couples d'images séparées par un très faible intervalle de temps (1jour) ; pour une telle période, il n'y a, en général pas de changement notable des propriétés de la surface, sauf dans le cas d'un événement exceptionnel (crue, avalanche, séisme, ...).

Nous avons d'abord observé (Fig. 3) le couple d'images acquises les 28 et 29 mai 1995. Les zones

apparaissant dans les couleurs rouge et verte, caractéristiques du changement, sont quasi inexistantes ; il y a donc très peu de changement, ce qui est normal sur un intervalle d'une journée. Les faibles surfaces rouges et vertes dans l'image (a) correspondent à des variations de radiométrie probablement dues à des effets d'atmosphère, ou plus sûrement à des défauts de calage. Les images (b) et (c) de la figure 3 présentent les mêmes changements ; les couples sont séparés par une journée (b : couple 28 mai 95 – 29 novembre 99 et c : couple 29 mai 95 – 29 novembre 99) ; mais les images constituant ces couples sont respectivement séparés de 3ans et 217 jours pour (b) et de 3 ans et 216 jours pour le couple (c).

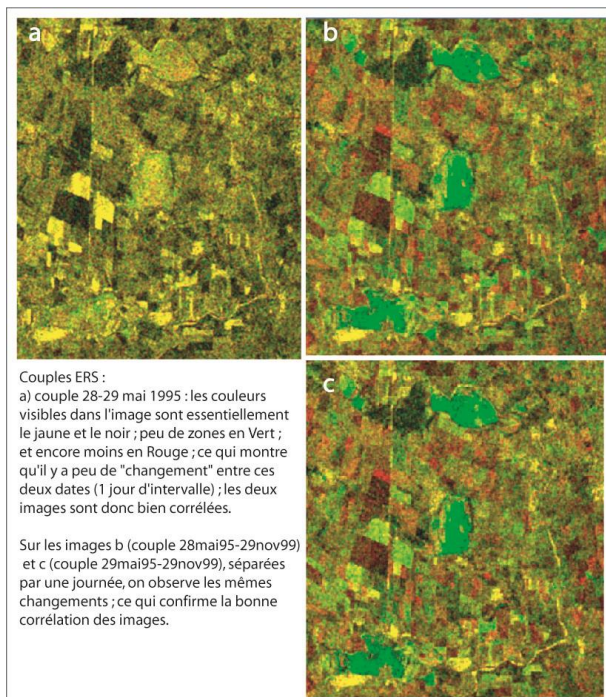


Fig. 3. La bonne corrélation des images

3.5. Les changements en zone de plaine

3.5.1. Les lacs

En 2004 :

La figure 4 montre la composition colorée associant les deux images acquises respectivement le 18 octobre et le 27 décembre 2004 ainsi que les précipitations des mois d'octobre et décembre 2004 à Buzău. Le premier lac (en haut de l'image) apparaît en rouge; cela signifie que le 18 octobre la surface du lac était recouverte d'eau (surface lisse); alors que le 27 décembre, le signal était fort (surface hétérogène à forte rétrodiffusion due soit à un vent fort – peu probable car les autres lacs ont une surface lisse- soit à la présence de végétation, soit un assèchement partiel du lac). La dernière

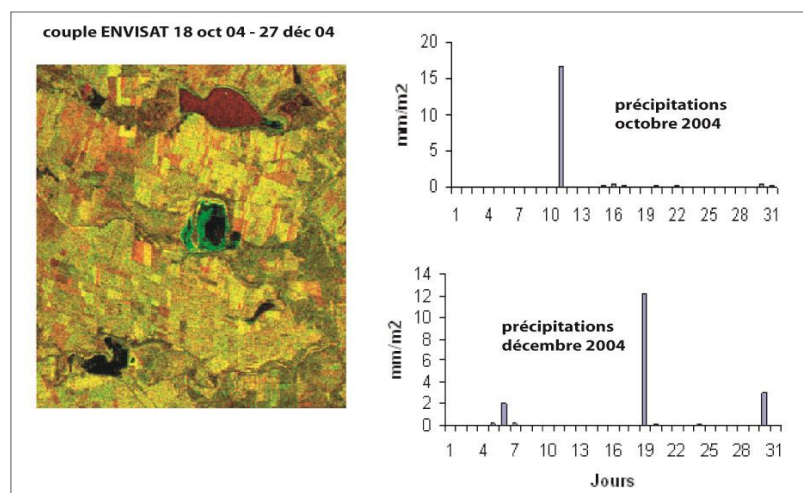


Fig. 4. Changements en 2004

hypothèse semble la plus probable si l'on considère l'extrémité droite du lac (apparaissant en noir, ce qui traduirait la présence d'eau produisant une surface lisse à signal rétrodiffusé faible) ; cela est d'ailleurs confirmé par le diagramme des précipitations (Fig. 5) qui montre une chute de la quantité d'eau tombée à partir de 1999 (période de déficit jusqu'en 2005).

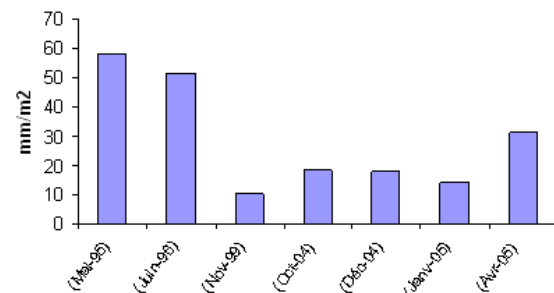


Fig. 5. Précipitation 1995-2005

En 2005 :

La figure 6 montre la composition colorée associant les deux images acquises respectivement le 31 janvier et le 11 avril 2005 (ainsi que les précipitations des mois de janvier et avril 2005 à Buzău. Le lac au nord de la zone semble avoir une surface pratiquement partagée en deux régions : une région nord en rouge, signe d'un signal fort en avril et faible en janvier ; et une région sud en jaune, signe d'un signal fort en janvier et également en avril. La région nord était donc recouverte d'eau en janvier, mais probablement pas en avril ; ceci est conforme aux graphiques des précipitations de décembre à avril : période pluvieuse en décembre et janvier (avec cependant un niveau faible de précipitations : 8 à 10 mm/m²) et relativement sèche en début d'avril (date de l'acquisition de l'image). La région sud traduit un signal fort en janvier et avril, cela est dû à une forte rugosité de la surface (pas d'eau, vent fort, végétation).

3.5.2. La végétation

La figure 7 montre les changements survenus dans la partie plaine (Nord Est de l'image) entre le 27 décembre 2004 et le 11 avril 2005 d'une part et entre le 31 janvier et 11 avril 2005 d'autre part. La date du 11 avril 2005 est donc commune aux deux couples.

Constatons dans un premier temps que les zones urbaines (Focșani et Rumniciu Sarat) n'ont pratiquement pas changé d'aspect ; une légère coloration rouge pour la composition colorée de gauche indique une rugosité (ou une humidité) plus forte pour le 11 avril 2005 (effet du vent, de la poussée de végétation ...).

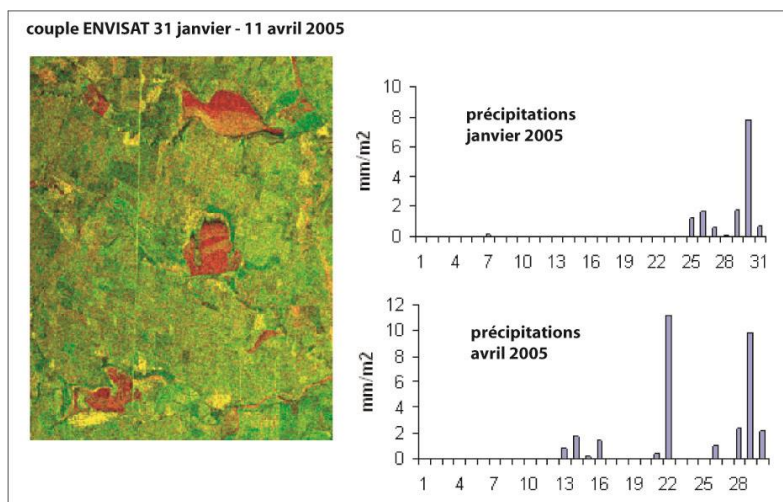


Fig. 6. Changements en 2005

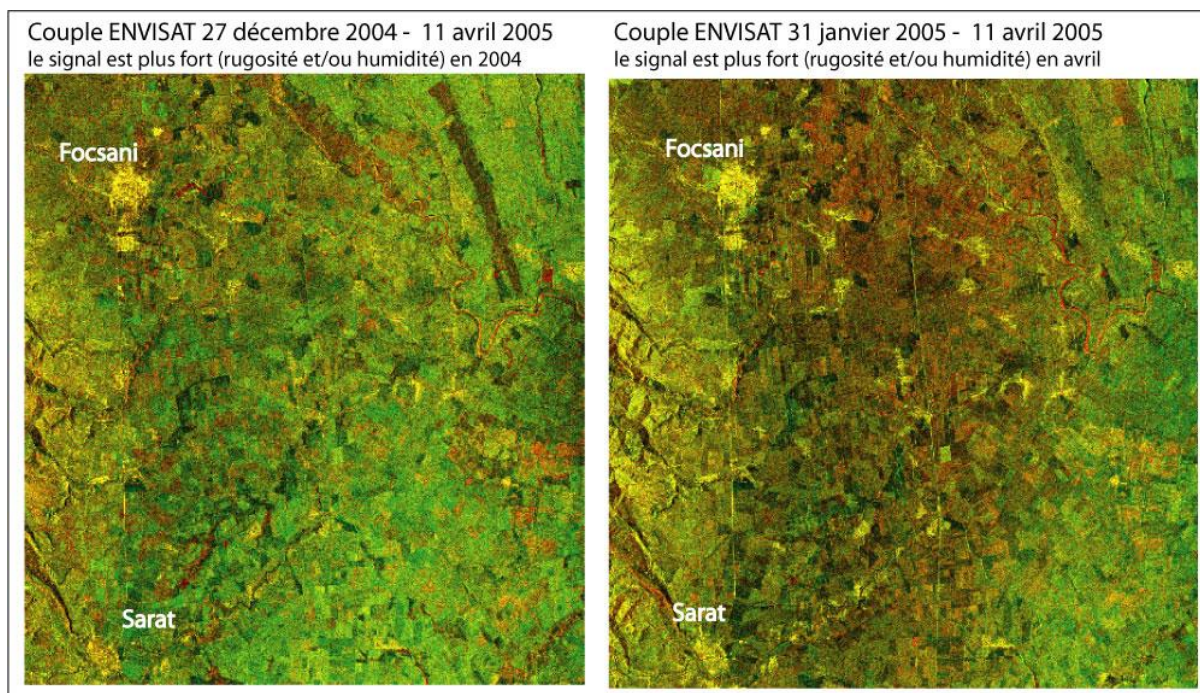


Fig. 7 : Changements en zone de plaine

Pour les parcelles agricoles, la situation est différente. On note un changement significatif entre les deux images de chacun des deux couples. Entre décembre 2004 et avril 2005, nous avons une diminution du signal (donc de la rugosité ou de l'humidité) ; par contre une augmentation du signal rétrodiffusé entre janvier 2005 et avril 2005.

Pour le lit de cours d'eau, par contre, nous notons une augmentation de signal dans les deux cas ; ce qui traduit une rugosité plus forte, autrement dit, probablement un dessèchement de surface (diminution de la quantité d'eau en surface). Ce résultat est conforme à celui obtenu pour les lacs. Il est donc plus plausible.

4. Premières conclusions

Les éléments principaux du paysage sont relativement bien distingués sur les images amplitude ; c'est le cas des zones urbaines, des parcelles et des lacs dans la zone plate ; les cours d'eau sont également bien distingués et la morphologie est bien décrite dans la partie accidentée ; ce qui prouve que l'image amplitude peut être concurrente aux images optiques malgré la présence du chatoiement (speckle).

Certains éléments demeurent néanmoins indiscernables à l'échelle des images et des déformations sur l'apparence des structures peuvent être observées, notamment dans la zone de relief, et donc fausser l'interprétation (apparition d'effet « cordon » par exemple). L'image radar d'amplitude, représente mieux la surface plate (plaine de Buzău) que la surface accidentée (collines subcarpatiques). Elle donne pour la région haute des distorsions qui sont proportionnelles aux degrés des pentes. Plus la pente est raide plus la distorsion est grande. Par contre, sur la zone des plaines, la distorsion est moins importante sauf pour la distorsion qui résulte des dimensions de résolution qui n'est pas exactement identique dans les deux directions.

Nous n'avons pas pu caler avec précision toutes les images de la série disponible et nous avons donc opéré un calage séparé des images provenant de chacun des deux satellites (ERS et ENVISAT) ; ceci a eu pour conséquence de réduire les dates d'analyse diachronique.

Nous avons observé que le signal rétrodiffusé est très cohérent dans la zone plate de l'image mais très bruité dans la zone de relief à cause des perturbations dues aux différentes structures et à des pentes différentes.

Signalons également que la lecture de l'image amplitude doit être réalisée avec plus de précaution que pour une image optique ; un signal de forte intensité qui n'est pas toujours lié à une forte rugosité de surface ou une humidité importante du sol. La cause peut être la pénétration de l'onde radar (diffusion du volume). Nous noterons enfin que les textures apparaissant dans une image radar sont dues à des surfaces rugueuses par rapport à la longueur d'onde (critère de Rayleigh) car divers niveaux de rétrodiffusion existent selon la nature de la surface et son orientation.

Au niveau des types d'images analysées, tout au moins pour notre zone d'étude, nous avons observé

que les données ENVISAT sont plus adaptées au suivi temporel grâce à leurs courts intervalles ; les images ERS auraient pu donner de meilleurs résultats ; mais leur intervalle temporel est tellement grand qu'ils génèrent des décorrélations temporelles surtout à cause des caractéristiques de la zone d'étude.

Pour synthétiser, un certain nombre de conclusions nous paraissent intéressantes à souligner de façon particulière :

- l'imagerie radar est certainement, parmi les images satellitaires, la plus adaptée dans la détection des changements de paysage, notamment les faibles déformations impossibles à détecter sur les images optiques ;

- plusieurs techniques de traitements des images radar existent ; mais il nous semble qu'aucune ne peut aboutir seule à une interprétation sûre des « phénomènes observés » ; ceci car les changements mis en évidence sur les produits dérivés peuvent avoir pour cause aussi bien des changements réels du paysage que des perturbations du signal enregistré : artefacts atmosphériques, conditions de prise de vue, etc. car la phase comme l'amplitude sont assez sensibles aux multiples sources du signal ;

- l'image radar, par le fait même qu'elle soit le produit d'une télédétection active, est complexe et son traitement l'est tout autant ;

- comme toutes les images ne permettent pas d'aboutir à des produits dérivés facilement interprétables, il est nécessaire de disposer d'une grande série d'images offrant ainsi un plus grand choix de sélection ;

- la nature de la zone enfin et le comportement du terrain influent grandement sur la complexité de la chaîne de traitement et la qualité des résultats pour un objectif thématique donné.

Le retour au terrain nous a permis de conforter la plupart des résultats avancés.

5. Remerciements

Nous remercions l'AUF pour le financement du projet de recherche partagée sur l'apport de l'imagerie satellitaire multi résolution dans le suivi des phénomènes de glissements de terrain en Roumanie qui a permis d'acquérir les images utilisées dans le présent travail.

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The Use of the Geomorphometric Method for the Assessment of the Giumalău Massif Drainage Development Stage

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Key words: drainage, hydrographic basins, base level, morphometric parameters

Cuvinte cheie: rețea de drenaj, bazine hidrografice, nivel de bază, parametri morfometrici

Rezumat: În Masivul Giumalău în condițiile unui substrat geologic, constituit în cea mai mare parte din roci metamorfice și al unor circumstanțe climatice care nu prezentau diferențieri spațiale importante, evoluția rețelei de drenaj a ținut seama, în primul rând, de nivelul de bază local dat de râurile Moldova și Bistrița. Demersul științific de față încearcă, prin metode morfometrice, să pună în evidență diferențierile existente între bazinele hidrografice de diferite ordine de mărime dar și nivelul până la care se resimt acestea, având în vedere faptul că există ipoteze care consideră că bazinul hidrografic al Moldovei s-a organizat mai repede decât bazinul Bistriței. Se constată o anumită diferențiere în ceea ce privește gradul de dezvoltare al rețelei de drenaj. Analiza parametrilor statistici indică un grad mai avansat de organizare la nivelul rețelei aferente bazinului Moldovei, nivelul de bază local fiind extrem de important pentru bazinele hidrografice de ordinele V și IV (sistem Strahler) și mai puțin semnificativ pentru bazinele hidrografice de ordinul III și mai mici.

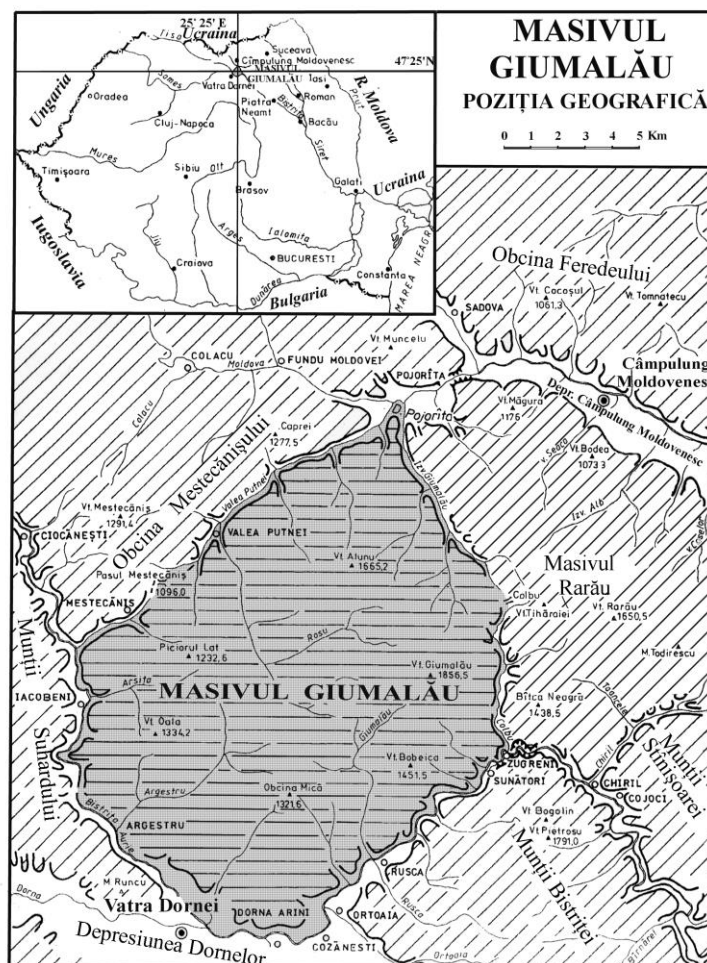


Fig. 1 Geographical location of Giumalău Massif

Drainage constitutes the main way through which eroded materials may be transported outside the Giumalău Massif perimeter. According to the geological time scale, material evacuation was tightly related to climatic manifestations, petrographic associations, vegetal composition and, last but not least, the development of the two important hydrographic basins: Moldova and Bistrița (Fig. 1).

The Giumalău Massif is part of the mezozoic crystalline area in the Oriental Carpathians, the present relief being the result of the interactions among multiple morphogenetic factors. The geological substratum is constituted mostly of metamorphic rocks (Balintoni, 1996), the morphotectonical and morphosculptural relief being differentiated into two tectonical compartments (eastern and western). The eastern one may be considered a horst raised by 500m above the western compartment. The second one, on the other hand, corresponding to the peak line Poiana Runcului-Mestecăniș, constitutes the watershed between Bistrița and Moldova rivers.

Considering the context, a great part of the morphosculptural and fluvial relief of the massif has to be related to the local base levels of Moldova's and Bistrița's thalwegs.

Thus, by means of morphometric methods, the present study attempts to emphasize the differences found among multiple size order basins and also the level up to which these are to be felt, considering the fact that Moldova's hydrographic basin was organized faster than Bistrița's one (Donisă, 1968, 1972, 1973).

The research started with the establishment of a hierarchical system for the considered territory drainage, with the use of the A.N.Strahler system (1952), which offers a logical support for the comparative geomorphological study of drainage basins in relation to different factors determining their evolution (Ohmori et al. 1993, Ichim et al., 1981, 1983, 1986, 1989, Bravard, 1998,). The Vth, IVth and IIIrd order basins were those used as a basis for the morphometric analysis (Zăvoianu, 1978, Grecu Florina, 1992, 2008).

The following have been identified: 49 IIIrd order hydrographic basins, totalizing an area of about 136.44 sqkm, which represent, within the massif limits, around 55% of the total area (Fig. 2.a.), 14 IVth order hydrographic basins and 3 Vth order hydrographic basins (Lesenciuc, 2000, 2004, 2006).

For the comparative analysis of the two drainage basins of Moldova and Bistrița, a series of variables associated to hierarchical orders, lengths, areas and frequencies have been determined: the number of Ist order rivers (magnitude), the drainage basin area (Sb, sqkm) and the minimum altitude in the basin (Hmin, M). To these, there may be added the concavity quotient of the longitudinal profiles of different sized order thalwegs.

From the total number of IIIrd order basins, 23 belong to Bistrița's hydrographic basin and 26 to Moldova's. As they have an approximately equal population in number, the statistical processing of these data was conducted according to the two greater hydrographic basins, in order to detect possible differences.

The analysis of the statistical parameters corresponding to IVth order hydrographic basins, grouped according to their affiliation to the two greater basins, was favoured by the equal number of 7 IVth order hydrographic basins belonging both to Moldova and Bistrița. The average basin area is of 11.11 sqkm for Bistrița and 7.50 sqkm for Moldova (Fig. 2.b.).

The comparative analysis of the average areas of differently ordered hydrographic basins emphasizes the following aspects:

- the Vth order basins of Moldova have an approximately fourth time larger average area than that of the Vth order basin belonging to Bistrița;
- a reversal occurs in the case of the IVth order basins: those belonging to Bistrița have a larger average area than those of Moldova;
- in the case of the IIIrd order basins, average areas corresponding both to Moldova and Bistrița are considerably similar (Fig. 3).

With respect to the areas of drainage basins, conclusions may be drawn on a general level. Differences in the average areas of the basins corresponding to Moldova and Bistrița are progressively diminishing from Vth to IIIrd order, which points to the idea that it is only in the case of hydrographic basins superior to the IIIrd order that the two local base levels influence Giumalău Massif from a morphosculptural point of view.

Thus, we may admit the fact that Moldova's IVth and Vth order hydrographic basins proved to have a more advanced drainage organization degree.

The comparative analysis of the two basins' minimum altitudes indicates a greater influence of

Moldova's and Bistrița's local base levels on Vth and IVth order basins and a decreasing one, to the point of disappearance, at the level of IIIrd order basins (Fig. 4).

The analysis of the local base average levels points to the fact that Moldova's Vth order drainage basins have their local base level by approximately 50m lower than the ones of Bistrița. Considering the base levels, Moldova's IVth order basins are

90m higher than the Bistrița's ones. In the case of IIIrd order basins, local base levels are situated at similar altitudes, with a plus of only 14m for Bistrița's basins. In conclusion, this variable also indicates that differences between the base levels of the drainage basins within the two main hydrographic basins have the tendency to decrease considerably from the Vth to the IIIrd degree.

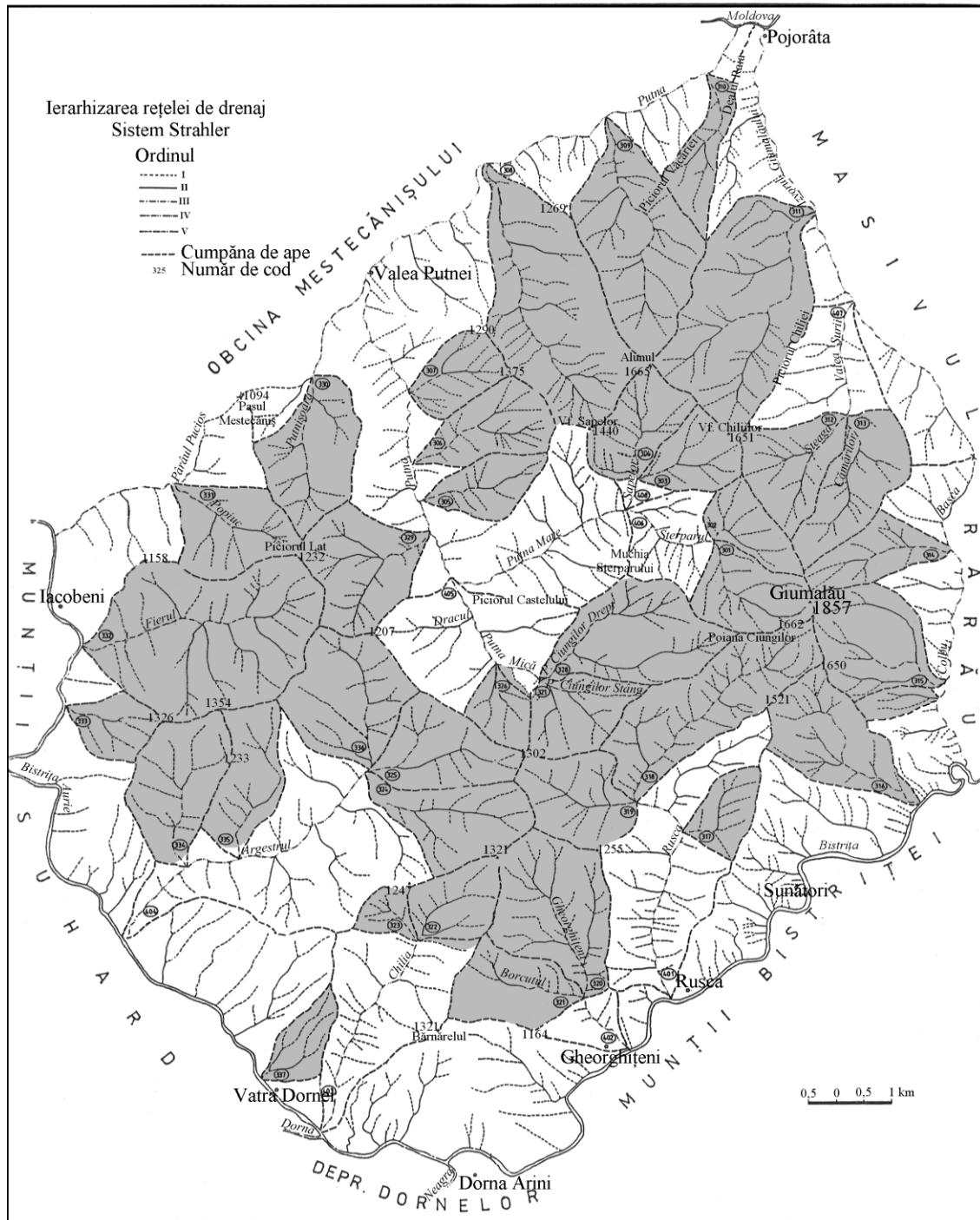


Fig. 2. a. Third order drainage basins distribution in the Giumalău Massif

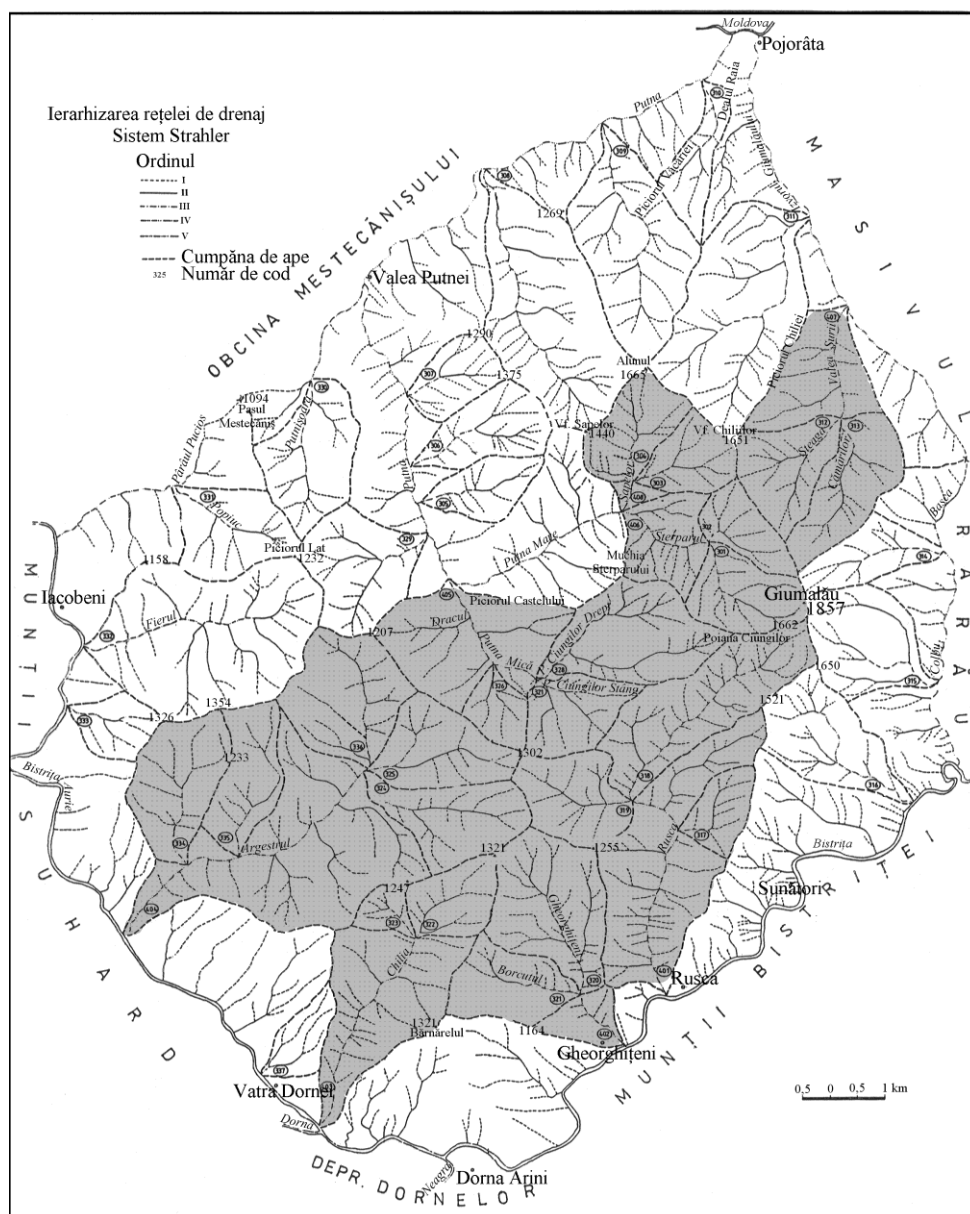


Fig. 2.b. Fourth order drainage basins distribution in the Giumalău Massif

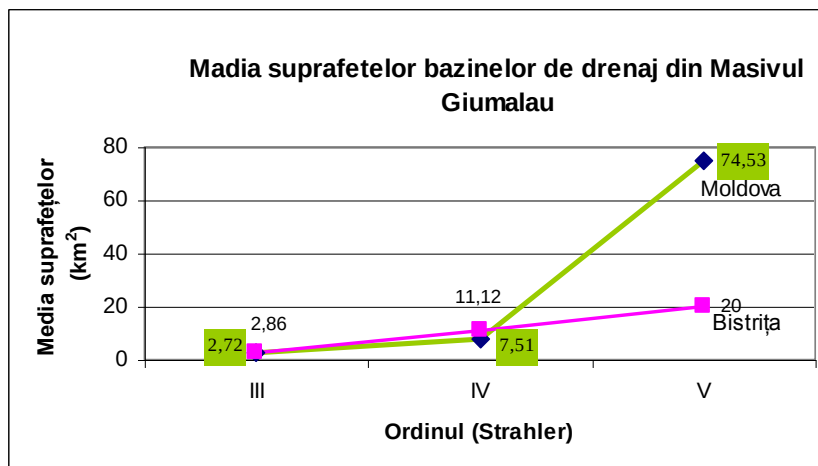


Fig. 3 Average areas of Moldova and Bistrița drainage basins

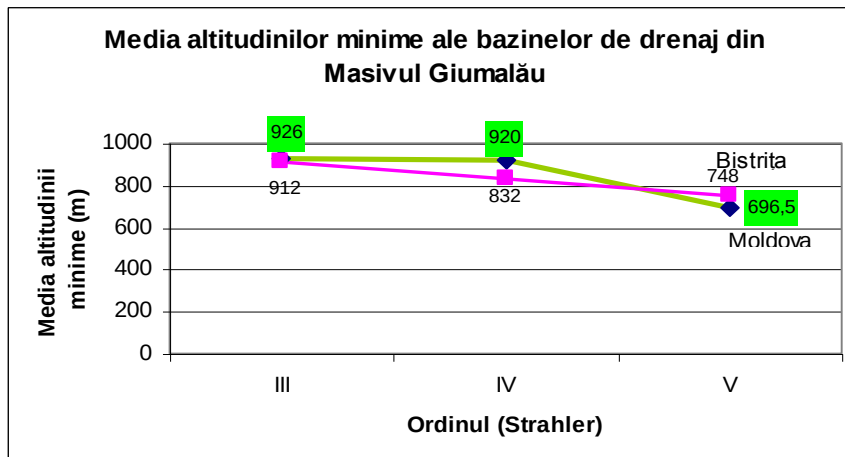


Fig. 4 Average minimum altitudes of Moldova and Bistrița drainage basins

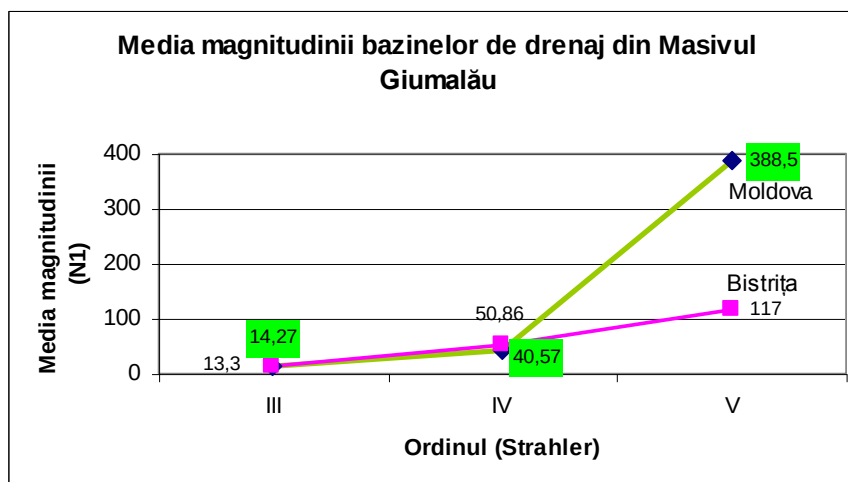


Fig. 5 Average magnitude Moldova and Bistrița drainage basins

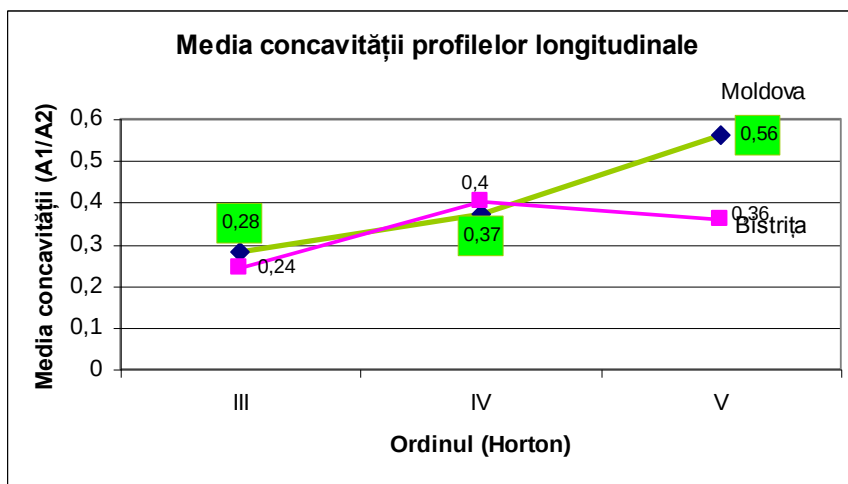


Fig. 6 Average concavity of longitudinal profiles in the Giumalău Massif

Regarding the magnitude (N1), significant differences between the two main hydrographic basins appear on the level of IVth and Vth order basins. Average values corresponding to IVth order basins are considerably greater for Moldova's basins, which may be explained by a much stronger

relief fragmentation in the area of Giumalău-Alunu peak line, to which may be added a local base level altitude situated approximately 50m lower than Bistrița's. Putna's Vth order drainage basin has a more advanced organization, covering more than half of the massif's total area. Bistrița's IVth order

drainage basins have average magnitude values superior to Moldova's, while on the level of the IIIrd order drainage basins, the magnitude is higher for those belonging to Moldova, although extremely similar to Bistrița's. The conclusion to be drawn once again is that magnitude differences diminish while the order decreases (Fig. 5).

The longitudinal profiles average concavity analysis demonstrates that the relationship between concavity and the local base level is obvious for all of the hydrographic basins categories (Vth, IVth and IIIrd orders), with a more advanced development for Bistrița's IVth order basins. The explanation lies in the fact that most of them are Bistrița's direct tributaries, while, on the level of Moldova's drainage basin, better developed IVth order tributaries (Putna and Izvorul Giumalăului) have at their turn local base levels situated at higher altitudes than Bistrița (Fig. 6).

We may thus conclude that Moldova's and Bistrița's local base levels represent two extremely important elements for the Vth and IVth order drainage basins evolution, this influence diminishing to the point of disappearance in the case of inferior order drainage basins.

At the level of IIIrd order drainage basins, distributed within multiple morphotectonical conditions, we may observe a significant reduction of the differences occurring between the two main drainage basins to which they belong.

In the context of a mostly metamorphic geological substratum and scarcely different climatic circumstances, drainage evolution was especially controlled by the two collectors local base levels.

A certain differentiation appears in the drainage development degree. The statistical parameters analysis indicates a higher degree of organization for Moldova's drainage basin. In these circumstances, we would be tempted to give the time factor an essential role, in accordance to some researchers opinion that Moldova drainage basin is older. But, taking into account a drainage network which does not go beyond the Vth order, we believe that its evolution degree is especially a result of the local base level established by these two main collectors – Moldova and Bistrița –, as well as a consequence of an adaptation to the area's tectonical and structural situation.

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Measurements of Various Erosion Processes in a Coastal Submediterranean Catchment (Southwest Slovenia)*

Matija ZORN

Key Words: geomorphology, soil erosion, rockwall retreat, gully erosion, chemical denudation, flysch, Istria, Slovenia

Abstract. Measurements of four different erosional-denudational processes were conducted between February 2005 and May 2006 in the Dragonja River Basin (Istrian Peninsula, southwest Slovenia): interrill soil erosion on three different land uses, rockwall retreat on steep flysch slopes, movements of debris in an erosion gully, and chemical denudation in the river basin.

Interrill soil erosion was measured on eight 1-m² erosion plots: on bare soil in a young olive grove (2 plots), in an overgrown meadow (2 plots) and in the forest (4 plots). The erosion plots in the forest were located on soil with two different slopes. The measurements (taken between May 2005 and April 2006) showed interrill soil erosion of 9,013 g/m² (90 t/ha; surface lowering of 8.54 mm) on bare soil in the olive grove with a slope of 5.5°; 168 g/m² (1.68 t/ha; surface lowering of 0.16 mm) in the overgrown meadow with a slope of 9.4°; 391 g/m² (3.91 t/ha; surface lowering of 0.37 mm) in the forest with a slope of 7.8°; and 415 g/m² (4.15 t/ha; surface lowering of 0.39 mm) in the forest with a slope of 21.4°.

The fastest erosion process was rockwall retreat in flysch badlands, with a specific erosion rate of 85 kg/m² per year, or almost 3.5 to 5 cm per year of rockwall retreat. Weekly measurements took place on four semi-open plots (open on the sides) ranging from 1.8 to 4.5 m² in size.

Movements of debris through an erosion gully were also measured in the same flysch badlands. A gully with a catchment area of approximately 1,000 m² and with an average slope of 46° was dammed. Almost 19 t of flysch material was accumulated behind the dam in twelve months.

A tipping bucket rain gauge was located next to the plots to monitor precipitation and intensity of erosive events.

Measurements of chemical denudation in the Dragonja River Basin were conducted monthly and showed a chemical denudation rate of 0.066 mm per year.

1. Introduction

In the river basins of the fluvial denudation relief (Gabrovec, Hrvatin, 2001), various erosion and denudation processes take place simultaneously; however, do we know at what speed they occur, and what the relationships between them are?

This is why we measured the speed of four erosion and denudation processes in the Dragonja River Basin (southwest Slovenia, the Slovenian part of the Istrian Peninsula; Fig. 1) in 2005 and 2006: interrill soil erosion on three different land uses, rockwall retreat on steep bare flysch slopes, movements of debris in an erosion gully, and chemical denudation in the river basin.

2. Measurements

Erosion and denudation processes are quantified either by means of measurements, or by various models (Stroosnijder, 2005). We decided to carry out measurements in the field. Advantages of research in the field are the possibility to conduct measurements at the proper scale, with realistic soil and plant characteristics and temporal changes in environmental variables (Stroosnijder, 2005), even though, at the same time, there is less control over dependent variables (such as weather conditions) than with measurements carried out in laboratories.

Measurement methods differ in scale, length of measurements, funding, and so on. Stroosnijder

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(2005) differentiates between five size categories for the study of erosion processes: (1) the point (1 m^2) scale for interrill erosion, (2) the plot ($< 100 \text{ m}^2$) for rill erosion, (3) the hillslope ($< 500 \text{ m}$) for sediment deposition, (4) the field ($< 1 \text{ ha}$) for channels, and (5) the small watershed ($< 50 \text{ ha}$) for spatial interaction effects. Stroosnijder (2005) also differentiates between two time criteria: one rainfall event and a yearly average. With regard to the time criteria, Loughran (1989) differentiates the rainfall event criterion and any other suitable time interval.

For measuring interrill soil erosion we, according to Stroosnijder (2005), used a “point” scale. After the same author we used “plot” scale for rockwall retreat, as well as for the movement of debris in erosion gullies. After Poesen, Torri, and Bunte (1994) we used micro erosion plots for measuring interrill soil erosion, mezzo erosion plots

for rockwall retreat, and macro erosion plots for debris movement in erosion gullies.

Measurements of soil erosion, rockwall retreat, and debris movement in erosion gullies were taken weekly, and water sampling for determining chemical denudation took place on a monthly basis. Especially for rockwall retreat, it is important that during the measuring period the average annual minimal air temperature in Koper, which is a few kilometers away, was 1.4°C lower, and the average winter minimal temperature 2.4°C lower than the long-term average (1961–1990; Klimatografija, 1995; Dnevne, 2006). For water erosion, it is important that during the measuring period the amount of precipitation at the nearby Portorož Airport was 85 mm or 8.2% higher than the long-term average (1991–1995) between May 2005 and April 2006 (Povzetki, 2007; Dnevne, 2007).



Fig. 1. The study area in southwest Slovenia

2.1. Interrill soil erosion

Soil erosion is the removal of soil particles and weathered debris by means of natural agents, in many cases accelerated by human activities (clear-cutting, overgrazing, and paths) and animals, and it is more intense than soil formation (Komac, Zorn, 2005; Zorn, Komac, 2005). It consists of interrill erosion (surface wash) and rill erosion. Our measurements were taken on closed erosion plots 1 m^2 in size that were used to measure interrill erosion (Fig. 2).

Measurements on erosion plots of similar size were, for example, taken in Spain (Dunjó, Pardini, Gispert, 2003; 2004; Usón, Ramos, 2001; Boix-Fayos et al., 2007).

The structure of erosion plots was found in a study by Vacca et al. (2000) and Ollesch and Vacca (2002). For the positioning of erosion plots in the landscape, works by Lal and Elliot (1994), and Dunjó, Pardini, and Gispert (2004) were used. Such small erosion plots are easily doubled and managed, but one must also keep in mind that such erosion plots usually overestimate actual erosion when extrapolated to larger spatial units because they do not take into account the sedimentation within larger spatial units (Collins, Walling, 2004).

The measurements took place in eight erosion plots. We placed two plots on bare soil in an olive grove with an average inclination of 5.5° , two in a meadow with an average inclination of 9.4° , and

four in a forest (two on a slope with an average inclination of 7.8° , and two on a slope with an average inclination of 21.4°).

As a rule, measurements were performed weekly. First, the volume of the collected surface runoff was determined, and then samples of water and detached soil mixture from the reservoirs were taken. The samples were dried at 105°C in the laboratory, where the concentration of undissolved particles was determined. A tipping bucket rain gauge equipped with a data logger was located next to the plots for monitoring precipitation and the intensity of erosive events (Zorn, 2008; Zorn, Petan, 2008).

Although our measurements are not long-term, it could still be seen that “larger” erosive events contribute a great portion to the total yearly soil loss. The biggest erosive rainfall event in the reference period occurred in the week from 5 August to 12 August 2005 (weekly erosivity was $1235.91 \text{ MJ}\cdot\text{mm}\cdot\text{ha}^{-1}\cdot\text{h}^{-1}$; on 11 August the maximum 30-minute precipitation was 42.8 mm, and daily erosivity was $1110.5 \text{ MJ}\cdot\text{mm}\cdot\text{ha}^{-1}\cdot\text{h}^{-1}$; the erosivity of precipitation was well above the August monthly average ($507.8 \text{ MJ}\cdot\text{mm}\cdot\text{ha}^{-1}\cdot\text{h}^{-1}$; Petkovšek, Mikoš, 2004) for the Dragonja River basin). During this week 30% of the total yearly soil loss was recorded on bare soil in an olive grove. In a meadow, 24% of the yearly soil loss was recorded. Because of the rainfall intercepted by treetops, the amount of soil loss in the forest was lower. In the forest with a lower inclination, 15% of the yearly soil loss was recorded during this week, and in the forest with a greater inclination the amount of soil loss did not stand out from other extremes. These

data indicate that the vegetation period has considerable importance for soil erosion (Zorn, 2008; Zorn, Petan, 2008).

On a yearly basis, 9 kg of soil/ m^2 is lost on bare soil, up to 170 g/ m^2 in meadows, up to 390 g/ m^2 in forests with a lower inclination, and 415 g/ m^2 in forests with a greater inclination. In the measurement period, the surface level on bare soil was reduced by almost one centimeter; in the forest this did not even amount to half a millimeter, and in the meadow it was negligible, amounting to one-fifth of a millimeter (Table 1). Table 5 includes measured values extrapolated to longer time periods, under the presumption that the conditions remained the same as during the measurement period. Under such a presumption, one may expect surface lowering by up to 9 cm in ten years, and up to a little less than one meter in one hundred years. Considering such data, the question arises whether the difference in the speed of soil erosion (or surface lowering) on various land uses is evident in the morphology of a landscape after a few centuries. The fact is that relief changes on farmland occur (see Komac, Zorn, 2005; Zorn, Komac, 2005). In some places in Slovenia, land has been farmed in the same place for over 1,000 years. For example, Mason (1995) mentions that near Adlešiči (southeast Slovenia) a “four-meter wide deposit” was transported from the upper to the lower terrace during the historical period.

In addition to the presented data, one must not forget that, according to Govers and Poesen (1988), interrill erosion contributes to only approximately 20% of all soil erosion.

Table 1. Interrill soil erosion and surface lowering under different land uses from 28 April 2005 to 26 April 2006 (Zorn, 2008)

	Bare soil			Meadow			Forest: lower inclination			Forest: greater inclination		
	Soil erosion		Surface lowering	Soil erosion		Surface lowering	Soil erosion		Surface lowering	Soil erosion		Surface lowering
	g/ m^2	kg/ha	mm	g/ m^2	kg/ha	mm	g/ m^2	kg/ha	mm	g/ m^2	kg/ha	mm
Avg. per week	173.34	1,733.35	0.16	3.23	32.34	0.003	7.52	75.22	0.01	7.98	79.78	0.01
Total (per year)	9,013.43	90,134.31	8.54	168.15	1,681.51	0.16	391.15	3,911.49	0.37	414.87	4,148.68	0.39

2.2. Sediment production in flysch rocks

Steep bare slopes (badlands; Fig. 4) are a morphogenetic peculiarity of the flysch part of the Istrian Peninsula. These include linear formations,

such as erosion gullies or torrent river beds (section 2.3), and surface formations that may appear as steep rock faces or ribbed relief on gentler slopes (Jurak, Fabić, 2000).

We used semi-open erosion plots in our measurements. They were semi-open because they were limited by the edge of the slope at the top of the plot (so they were closed on the upper side), and

they were open on the sides (Fig. 5). We set up four erosion plots. The sizes of the contributing areas were from 1.8 to 4.5 m².



Fig. 2. Erosion plot on bare land in an olive grove with collected weekly runoff and eroded material (7 April 2005–14 April 2005; photo: Matija Zorn)

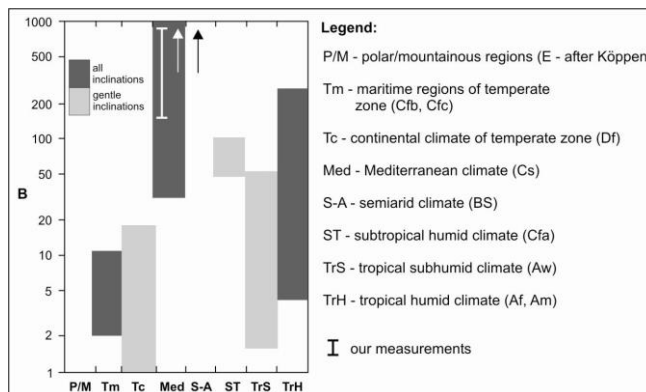


Fig. 3. Placement of measurements in a Mediterranean climate, and comparison with other measurements in other types of climate (1 B (Bubnoff) = 1 mm/1,000 years = 1 m³/km²·per year). Erosion data for various climate types was gathered by Saunders and Young (1983; Young & Saunders 1986)



Fig. 4. The badlands where the measurements were taken (photo: Matija Zorn)



Fig. 5. Semi-open erosion plots for flysch rockwall retreat measurements (photo: Matija Zorn)

Sediment production amounted to approximately 85 kg/m² per year between 28 April 2005 and 26 April 2006 (Table 2, Fig. 6), which means that the slope is retreating from approximately 35 mm per year (taking into account the bulk density of flysch rocks: 2,300 kg/m³ according to Miščević, Števančić, Štambuk-Cvitanović, 2008) to almost 50 mm per year (taking into account the bulk density of flysch rocks: 1,680–1,700 kg/m³ according to Petkovšek, Klopčič, Majes, 2008). The extrapolation of these measurements to longer time periods indicates that slopes may retreat by approximately 0.35 to 0.50 m in ten years, or 35 to 50 m in one hundred years (Table 5; Zorn, Mikoš, 2008).

Considering these data as well as soil erosion, the issue of the role of badlands in recent morphodynamics raises. If we assume that badlands in Mediterranean landscapes were formed during the historical period (Poesen, Hooke, 1997), then the Dragonja River Valley has expanded by a few dozen meters in the badlands in the period of a few hundred years. Major transformations (mostly deforestation) in the Dragonja River Basin have been recorded since the middle of the fifteenth century. If we are to believe De Ploey (1992) that badlands in southern Europe are between 2,700 and 40,000 years old (Poesen, Hooke, 1997), then they appeared during even more extreme climate

conditions and their role in the formation of valleys in Mediterranean landscapes is underrated in

comparison to the combination of stream erosion and neotectonic activity.

Table 2. Sediment production and rockwall retreat on steep flysch slopes from 28 April 2005 to 26 April 2006 (Zorn 2008)

	Sediment production		Rockwall retreat
	kg/m ²	kg/ha	mm
Average per week	1.62	16,203.47	0.70–0.95
Total (per year)	84.26	842,580.20	36.63–49.22

The speed of rockwall retreat clearly indicates the great erodibility of flysch rocks, which is also evident in Figure 6.

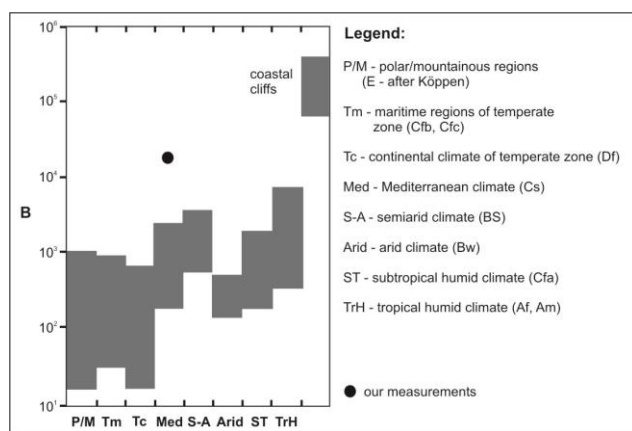


Fig. 6. The placement of measurements in a Mediterranean climate and comparison with measurements in other climate types (1 B (Bubnoff) = 1 mm/1,000 years = 1 m³/km² per year).

Erosion data for various climate types were gathered by Saunders and Young (1983; Young, Saunders, 1986)

During the measurement period, we noticed three peaks in sediment production (Fig. 7). The primary peak is connected with the alternation of temperatures above and below freezing during the daytime in the colder part of the year. The secondary and tertiary peaks are connected with heavier precipitation in the spring and summer. The same conclusions were reached in the Pyrenees (Spain) (Regüés, Pardini, Gallart, 1995; Regüés, Guàrdia, Gallart, 2000); namely, that heavy weathering is characteristic for the part of the year with below-freezing temperatures, and that, for the warmer part of the year rain (water) erosion is more characteristic. They claim that the amount of sediment produced depends on the speed of weathering (Regüés, Pardini, Gallart 1995). Regüés, Guàrdia, and Gallart (2000) discovered that sediment production caused by intensive weathering in the winter is “about two orders of magnitude greater” than sediments produced due to rainwater erosion (Fig. 8).

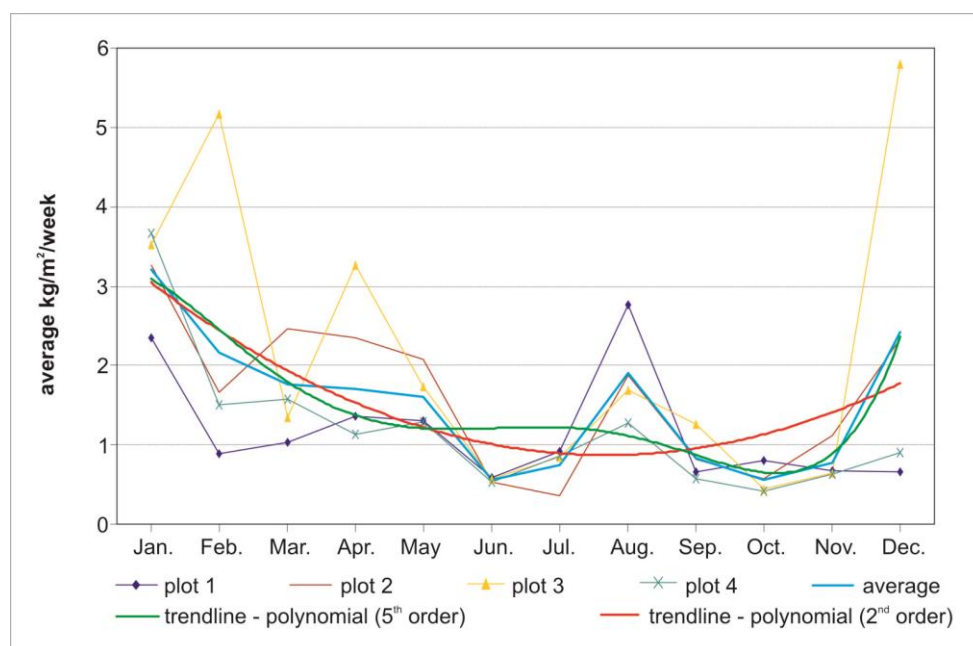


Fig. 7. Sediment production in flysch rocks by month

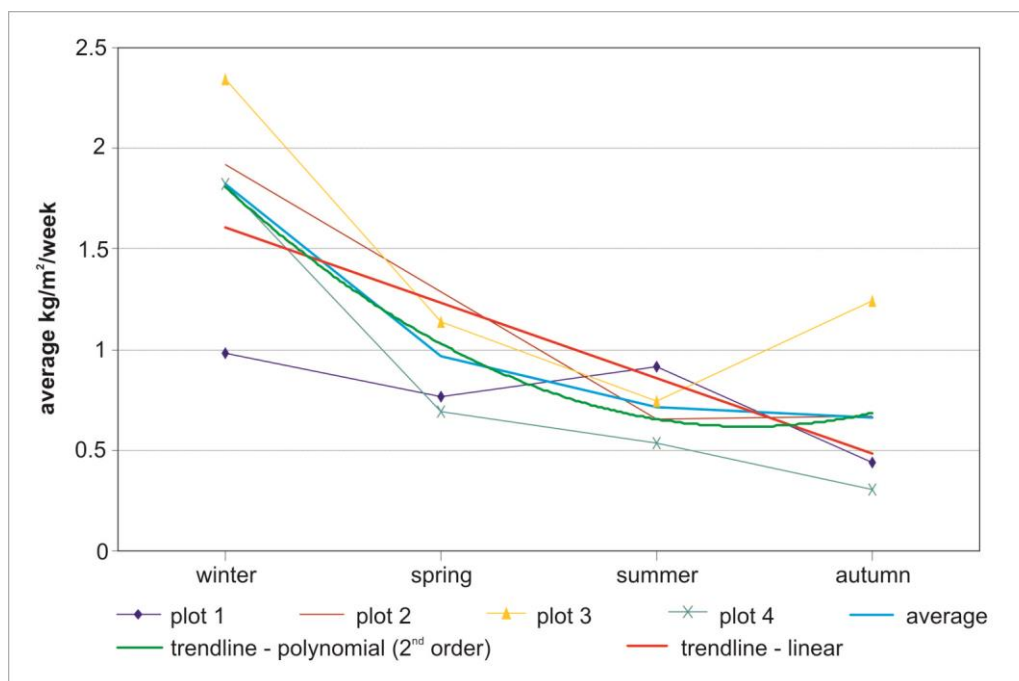


Fig. 8. Sediment production in flysch rocks by season

The data by season clearly show the trend (Fig. 8) of the decrease in sediment production from winter to fall. Sediment production is most intense in winter, followed by a nearly 50% decrease in spring, and then an almost even amount of sediment production in summer and fall. This indicates a close connection between sediment production and the cold part of the year. The lower degree of sediment production in summer and fall is mostly connected to the lower intensity of weathering, which causes less material to become unstable. Because there is less unstable material available, even heavy precipitation does not cause as much material to be released as frost weathering in winter.

2.3. Movement of debris in erosion gullies

Steep bare slopes include not only flysch rockwalls (with a 70–90° inclination; section 2.2) but also slopes affected by erosion rills and gullies. Erosion gullies are less steep than rockwalls (40–60° inclination) (Radinja, 1973).

Erosion gullies are formed by channelization of surface streams, mostly in places where there are already some natural outflow lines, as well as along various anthropogenic formations (such as excavations, furrows, plot boundaries, paths, and roads). In contrast to interrill and rill erosion, in which the material is for the most part deposited on the slope itself or at its bottom, gully erosion offers

more chances for eroded material to reach streams in the valley bottom.



Fig. 9. Dam in the erosion gully (photo: Matija Zorn)

According to Poesen and Hooke (1997), there is neither a standardized methodology nor a universal model for measuring gully erosion. Problems

already appear in determining the size of erosion gullies because they are “partly subjective,” say Wainwright and Thornes (2004), and most frequently their lower boundary is considered the surface of the gully’s cross-section $> 900 \text{ cm}^2$ ($30 \times 30 \text{ cm}$).

We used a closed erosion plot to measure the movement of debris in the erosion gully because we used a dam to close off the entire catchment, which covered an area of 0.1 ha.

Between 28 April 2005 and 26 April 2006, almost 19 t of flysch debris accumulated behind the dam, which makes the sediment delivery almost 15 kg/m^2 per year (Table 3). This means that slopes in the gully retreated by approximately 6.3 to 8.5 mm in one year. Extrapolation to one hundred years indicates a slope retreat in the gully by just less than one meter (Table 5).

Table 3. Movements of flysch debris in the erosion gully between 28 April 2005 and 26 April 2006 (Zorn, 2008)

	Sediment delivery		Slope retreat
	kg/m^2	kg/ha	mm
Weekly average	0.28	2738.86	0.12–0.16
Total (in a year)	14.46	145,159.70	6.31–8.48

The effect of extreme events on erosion processes turned out to be the greatest in the movement of debris through erosion gullies. In the week between 19 January 2006 and 26 January 2006, as much as 52% of the year’s amount of debris was deposited behind the dam, and an additional 30% was deposited within seven weeks, with more than 3% of the yearly amount of debris caught. In the remaining 44 weeks, only 18% of

debris was caught behind the dam. The peaks are a result of dry rock flows that are triggered in the gully if there is enough debris in the gully, and if there are strong winds that completely dry out the debris to a certain depth. Saturated flysch debris with many clay particles is not as mobile, and moves only in times of heavier precipitation.

The largest amount of debris was moved through the gully in the first three months of the year, and the secondary peak was in August (Fig. 10). August was the month with the highest erosivity of precipitation; however, it lags behind the first three months of the year regarding the amount of the aforementioned rock flows in the week prior to 26 January 2006. These flows also appeared in February and March, but there was not as much debris in the gully.

The effect of the amount of debris on the movement in the gully is shown by a comparison between the week prior to 23 March 2006 and the week prior to 12 August 2005. In the week of March, the erosivity of precipitation amounted to $100 \text{ MJ}\cdot\text{mm}\cdot\text{ha}^{-1}\cdot\text{year}^{-1}$, and 3.34% of a yearly value of debris was moved. In the week of August, the erosivity of precipitation was no less than $1,235.91 \text{ MJ}\cdot\text{mm}\cdot\text{ha}^{-1}\cdot\text{year}^{-1}$, and twice as much debris was moved. In the first case, there was much debris in the gully because many alternation between above-freezing and below-freezing temperatures caused significant sediment production on its slopes. Towards the summer, sediment production on the slopes decreased, and there was less debris in the gully from week to week. The lack of debris in the gully was even more evident in the fall months, when the least amount of debris was moved despite fall precipitation (especially in November; Fig. 11).

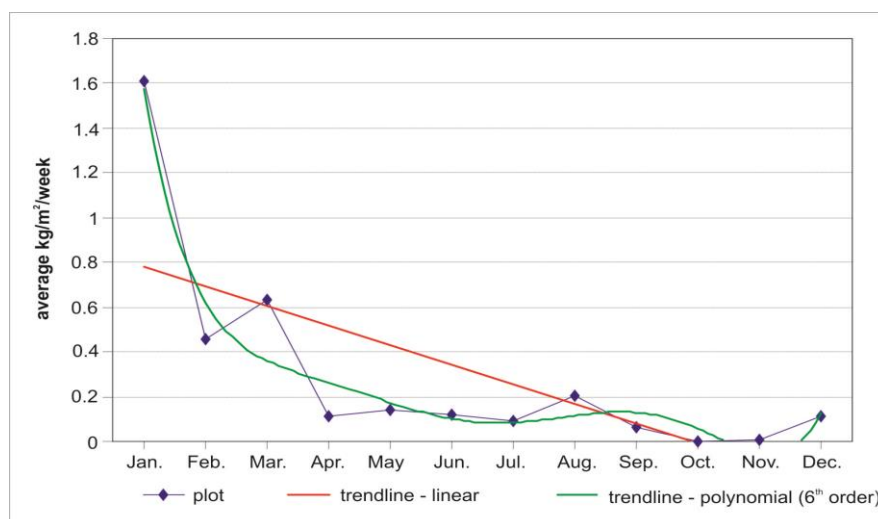


Fig. 10. Monthly movements of flysch rocks in the erosion gully

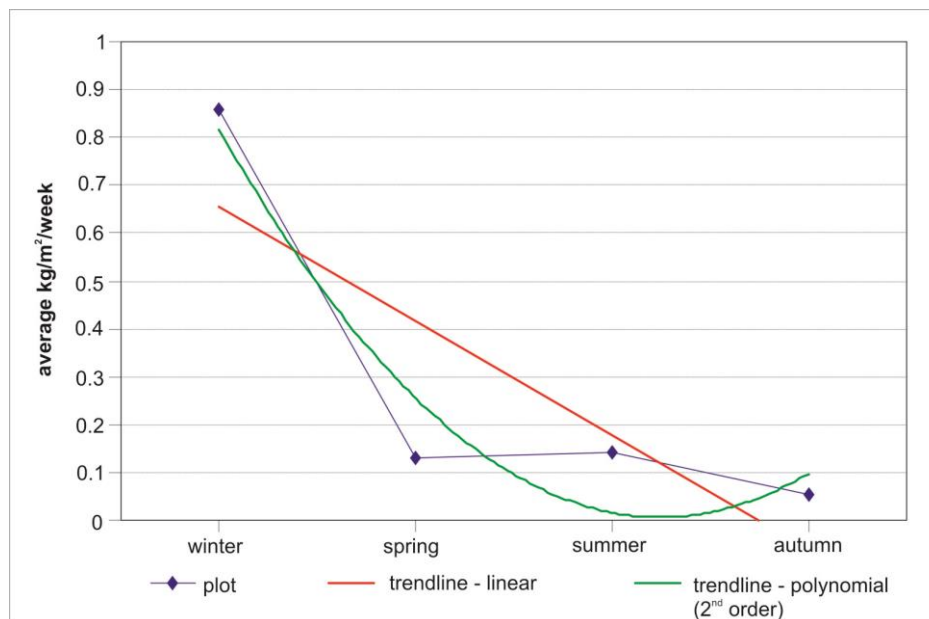


Fig. 11. Seasonal movements of flysch rocks in the erosion gully

Table 4. The speed of various erosion and denudation processes and the ratio between them from 28 April 2005 to 26 April 2006 (Zorn, 2008)

	Rockwall retreat of bare steep flysch slope	Movement of debris in erosion gully	Soil erosion on bare soil	Soil erosion in meadow	Soil erosion in forest (lower inclination)	Soil erosion in forest (greater inclination)	Chemical denudation
Specific sediment production (g/m ² per year)	84,258.02	14,455.27	9,013.43	168.15	391.15	414.8	113.32
Ratio to bare steep flysch slope	1.00	0.17	0.11	0.002	0.005	0.005	0.001
Ratio to erosion gully on bare steep flysch slope	5.83	1.00	0.62	0.01	0.03	0.03	0.008
Ratio to soil erosion on bare soil	9.35	1.60	1.00	0.02	0.04	0.046	0.013
Ratio to soil erosion in meadow	501.09	85.97	53.60	1.00	2.33	2.47	0.67
Ratio to soil erosion in forest (lower inclination)	215.41	36.96	23.04	0.43	1.00	1.06	0.29
Ratio to soil erosion in forest (greater inclination)	203.10	34.84	21.73	0.41	0.94	1.00	0.27
Ratio to chemical denudation	743.53	127.56	79.54	1.48	3.45	3.66	1.00

2.4. Chemical denudation

Most frequently, chemical denudation is measured in karst relief but can also be measured in other places if the rocks contain a high level of carbonates. The flysch rocks of Istria contain up to 75% carbonates (Magdalenčić, 1972).

In determining chemical denudation in the Dragonja River Basin, we measured water hardness

at the water gauging station below Kaštel once a month, a few kilometers from the river's outfall into the Adriatic Sea. We measured the amount of dissolved matter, especially calcium (Ca^{2+}) and magnesium (Mg^{2+}) ions.

Some earlier authors (Gams, 1974; 2003) observed that the water hardness in rivers on flysch rocks in the Slovenian part of Istria is relatively high

compared to the water hardness in karst areas. This is also true of flysch rocks in the Slovenian Julian Alps (northwest Slovenia) (Komac, 2001).

Our results indicate that chemical denudation in the Dragonja River Basin causes surface lowering at a rate of 66 mm per 1,000 years. This is slightly more than what Gams (1974; 2003) stated when he estimated 62 mm per 1,000 years.

2.5. Comparison of Measurements

Various erosion and denudation processes occur in basins simultaneously but their speed varies

significantly. The measurements described enable us to compare the speeds of some of them.

The fastest are the processes on steep bare flysch slopes, followed by the movement of debris in erosion gullies, which is almost six times slower. Interrill erosion on bare flysch soil is almost nine times slower. Soil erosion in the forest is over 200 times slower and in the meadow over 500 times slower than sediment production in flysch rocks on steep slopes. The slowest is chemical denudation, which is 744 times slower than sediment production on steep flysch slopes (Table 4, Fig. 12).

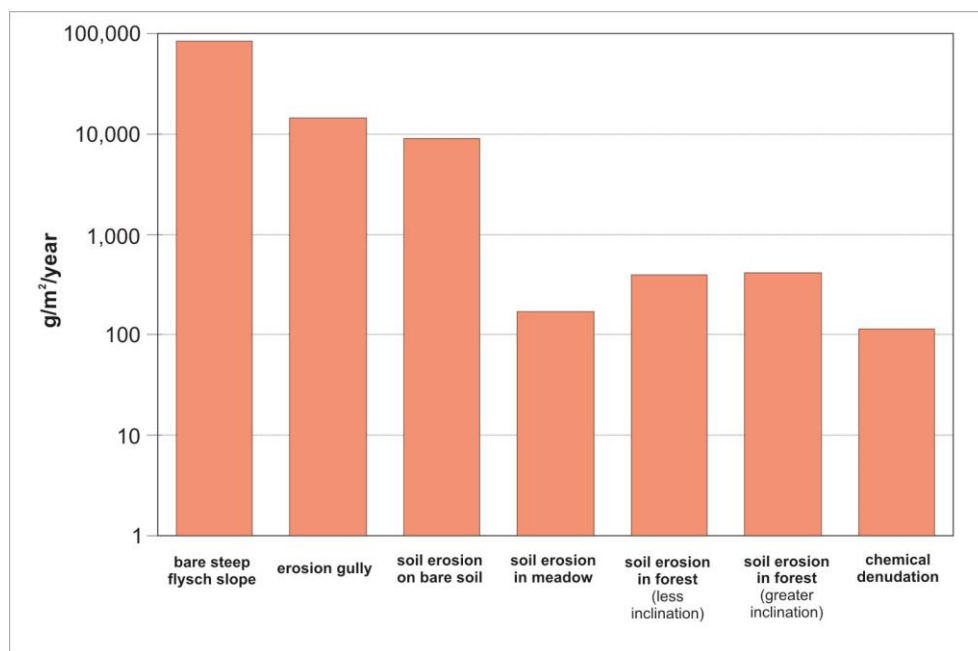


Fig. 12. Speed of various erosion and denudation processes between 28 April 2005 and 26 April 2006 (Zorn, 2008)

Table 5. Surface lowering/rockwall retreat in various time spans, assuming that the conditions are the same as at the time of the measurements (28 April 2005–26 April 2006) (Zorn, 2008)

In ... years	Bare steep flysch slopes	Erosion gully	Bare soil	Meadow	Forest: lower inclination	Forest: greater inclination	Chemical denudation
	Slope retreat	Slope retreat	Surface lowering	Surface lowering	Surface lowering	Surface lowering	Surface lowering
	m	m	m	m	m	m	m
1	0.049	0.009	0.0085	0.0002	0.0004	0.0004	0.00007
10	0.49	0.09	0.085	0.002	0.004	0.004	0.0007
100	4.92	0.85	0.854	0.016	0.037	0.039	0.007
1000	49.22	8.48	8.54	0.16	0.37	0.39	0.066

3. Conclusion

The measurements presented were taken during a period that was above average concerning the amount of precipitation and low temperatures (and

the transition from below-freezing to above-freezing temperatures). These measurements therefore indicate above-average sediment production. However, climate deviations were not great, and they are true for only the past few decades. Colder

periods were already characteristic for previous centuries (e.g., the Little Ice Age). On the basis of the measurements presented, it is therefore easy to claim that denudation and erosion processes in Mediterranean flysch landscapes are rapid. The slopes of badlands are retreating by a few

centimetres per year, and soil erosion is causing unprotected slopes to lower up to one centimeter a year. Due to this intensity, erosion and denudation processes are classified among significant morphogenetic factors in Mediterranean flysch regions.

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Records of Climate Change over the Late-Holocene in the Danube Delta Coastal Dune System

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Key words: aeolian activity, Danube delta, parabolic dunes, Optical Stimulated Luminescence (OSL), dendrochronology

Abstract: The largest dunefields in the Danube delta developed during the Late-Holocene on the Caraorman and Letea prograded barriers. The sand dunes in the Danube delta are one of the most pristine coastal dunes in Europe. Our research focused upon the reconstruction of aeolian landforms in the Danube delta.

Field based investigations of the modern aeolian morphology revealed the succession of different aeolian activity phases shaping the modern aeolian landforms. One phase of aeolian modification of the landforms occurred during conditions of generally low rugosity (bare and grassy areas) while another one followed when the development of aeolian landforms was constrained by the emplacement of woody vegetation. The highest altitudes of the prograded barrier at Letea correspond to the accretion fronts developed in association with the forest.

Mapping of the morphologic features of the Letea complex ridge plain shows the discordant superposition of the aeolian landforms over the marine wave-laid sediments (beach ridges). The relationship between the resultant migration direction of the dunes (RMDD) and the underlying beach ridge orientation is an important controlling factor on the development of the aeolian morphology. Elongated parabolic dunes occur where the RMDD is concordant with the subjacent beach ridge alignment.

A combination of dendrochronological dates and optically stimulated luminescence (OSL) ages were used in order to develop a chronology of aeolian activity in the Danube delta during the Late-Holocene. The preliminary OSL ages indicate that the interplay between sea-level, sediment availability and climate variability over the last 2400 yrs resulted in a dunefield pattern representing several dune generations.

Our OSL ages, together with the ages produced by Giosan et al. (2006) advocate for a new hypothesis regarding Letea complex ridge plain development according to which at least three phases with different rates of progradation occurred: (1) rapid rates at the beginning, since 3640 ± 460 yr B.P. to 2300 ± 480 yr B.P., (2) slow rates since 2300 ± 480 yr B.P. to 900-1200 yr B.P. and (3) rapid rates since 900-1200 yr B.P. until the initiation of the Chilia secondary delta.

Introduction

The open coast Danube delta started to develop 5210 ± 280 yrs. B.P. (Giosan et al., 2006), leading to the development of a 30 km wide deltaic plain. Delta progradation occurred at different rates resulting in distinct morphological settings. One of the most significant morphological characteristics of the Danube delta is represented by the extensive dunefields from Letea and Caraorman prograded barriers. These dunes represent the largest aeolian landforms (aeolian sand volumes) from the Black Sea basin. They stand for one of the most pristine coastal dunes in Europe. Despite their potential role in the Danube delta development and their capacity to deliver information about the paleoenvironmental changes, these dunes, like all the coastal dunes from the Black Sea basin, remain little studied, representing thus a missing piece of the general picture of Holocene aeolian activity in Europe.

This is the first study aiming to determine the emplacement time of dune fields from the Danube delta and to give a first description of their internal structure, morphology and development pattern. The dunefields developed on the Letea and Caraorman prograded barrier represent geomorphic features that encompass high resolution records of coastal and aeolian processes involved in their development. Our data together with those produced elsewhere across Europe (Clemmensen et al., 1996, Clemmensen et al., 2001, Orford et al., 2000, Wilson et al., 2001, Wilson, P., 2002, Clarke et al., 2002, Clarke and Rendell, 2006, Buynevich et al., 2007, Clemmensen et al., 2007) contribute to the building of a general picture of Late Holocene aeolian activity in Europe. Understanding the effects of environmental and anthropogenic changes on coastal landscapes is particularly important in a current regime of rapid shifts in climate, sea level, and sediment supply,

combined with ever increasing population pressures along sandy coasts (Buynevich et al., 2007). Furthermore, this information could be used for predicting the possible effects which changes in sea level and wind/wave climate may have on coastal environments during the next century (Pye and Neal, 1993).

A combination of optically stimulated luminescence (OSL) ages, dendrochronological dates as well as field based morphological investigation and maps interpretation were used in order to develop a preliminary chronology of the late Holocene aeolian activity in the Danube delta.

Methodology

Two fieldwork campaigns were undertaken in autumn 2006 and spring 2007 when geomorphological observations and sand samplings for OSL dating were carried out on the Letea and Caraorman prograded barriers. Topographic measurement were performed in order to obtain the Digital Elevation Model (DEM) of morphodynamically representative parabolic dunes using a Sokkia 610 total station.

Dune age was determined using optical stimulated luminescence (OSL). Optical dating is used to determine the period of time elapsed since the mineral grains were last exposed to daylight. Aeolian sands are ideal materials for the application of OSL dating technique because quartz grains are usually well exposed to light during aeolian transport (Wintle, 1993). The time since last exposure to the daylight is calculated by dividing the apparent dose (Gy) absorbed from naturally occurring ionizing radiation by the dose rate (Gy/ka). In our study, clean sand size grains of quartz were obtained after sieving and chemical treatment (10% HCl for carbonate removal, H₂O₂ for organic fraction removal). The 180-212 µm quartz fraction was separated of heavy minerals and feldspars using Sodium Polytungstate (with a density 2.7 g/cm³, respectively 2.62 g/cm³). The new obtained quartz fraction was then treated with HFl for the external layer removal, than re-sieved. The 150-180 µm was then used for aliquots preparation under subdued orange lighting. The radiation measurement was performed with a TL/OSL Risø reader equipped with an EMI 9635QA photomultiplier and 2 Hoya U-340, 3 mm thick filters. Single Aliquot Regeneration (SAR) dose protocol (Duller, 2004) was used to estimate the

radiation dose that the samples had absorbed during burial. This protocol makes repeated OSL measurements of each sub-sample, or aliquot, so that the growth of the luminescence signal as a function of laboratory radiation dose can be determined. The importance of SAR procedure resides in OSL signal sensitivity changes assessment and addressing this issue.

Within the SAR sequence, all OSL measurements were made while holding the sample at 125°C. The natural and regeneration signals were measured following a preheat step of 220°C for 10 seconds, and the test dose following heating to 160°C. To test the appropriateness of this SAR protocol to these samples, between 2 and 6 aliquots of each sample were bleached in the laboratory (using two sets of 100s exposures to blue diodes) and then given a known laboratory radiation dose. Additionally, 20 aliquots of Letea 5 were exposed to the sunlight and then the dose recovery test was performed. The ratio of the given laboratory dose to that which was measured varied from 0.96 to 1.03, demonstrating that SAR is appropriate for these samples.

In order to build a long dendrochronological series living old oak trees were chosen in accordance with dendrochronological principles (Fritts, 1976; Cook and Kairiukstis 1990; Popa 2004a). Because of protective status of forest only one core have been extracted from each tree with an increment borer at breast height. All samples were dried out and sanded to improve the visibility of tree-ring borders. The LINTAB equipment and TSAP software were used for measuring the annual ring-widths with a precision of 0.01 mm, as well as for cross-dating the growth series by graphical comparison (Rinntech, 2005). The results were checked for missing ring and dating error using the COFECHA software through the analysis of the correlation on intercalated subperiods (Holmes, 1983; Grissino-Mayer, 1997). The final dataset comprises 55 individual series (25 from Caraorman and 30 series from Letea). The growth series were standardized in order to eliminate the non-climatic trend and to maximize the climatic information from the individual series. A double standardization was applied: first a rigid cubic spline with a 50% frequency response cutoff width equal to 200 years to remove the age effect and a second, more flexible, cubic spline of 10 years to preserve only the high frequencies signal.

All detrended series were averaged to chronology using biweight robust mean (Cook, 1985).

Regional settings

Letea and Caraorman dunefields developed in the central part of the Danube delta, at about 25 km landward (Westward) from the modern shoreline. The Danube delta dunefields currently evolve under a temperate dry climate, with average precipitation amount of about 300-350 mm/year which renders it the driest region in Romania with 104 days of drought and 200.3 dry days at Sulina (Barbu and Popa, 2002). A recent study undertaken by Bălceanu et al. (2008) reveals a tendency toward a drier climate with a decrease of the mean annual precipitation recorded at the meteorological stations from the Danube delta during the last four decades: -2.4 mm/yr at Sulina, -3.1 mm/yr at Gorgova and -3.7 mm/yr at Sfântu Gheorghe. The mean multiannual temperature is 11°C with sharp differences between the seasons (January: -0.2 °C; August: 21.8 °C). The wind climate is macroenergetic (63.03 vu according to Fryberger technique, 1978) and the prevailing winds are from

north and display an accute bimodal distribution: NW-NE (Preoteasa and Vespremeanu-Stroe, 2004). As a consequence, the aeolian morphology is mainly represented by parabolic dunes, elongated parabolic dunes and nebckha dunes.

Danube delta developed under the alternating control of fluvial and marine factors. Since the beginning of delta development, Danube sediment discharge fluctuated in relation with the climate variability (Giosan et al., 2006), but generally recorded the same order of values as in the period before damming: 6.8×10^{10} kg/yr (Almazov et al., 1963). Black Sea is a tideless basin and the significant wave height measures 0.9m (Vespremeanu-Stroe, 2004). The modern longshore sediment drift values were assessed through sedimentary budget method - computations of reworked sediment volumes (0.7×10^6 m³/yr - Panin, 1972, 1984) and by modeling the nearshore wave transformation (0.85×10^6 m³/yr - Giosan et al., 1999; 1×10^6 m³/yr - Vespremeanu-Stroe, 2004).

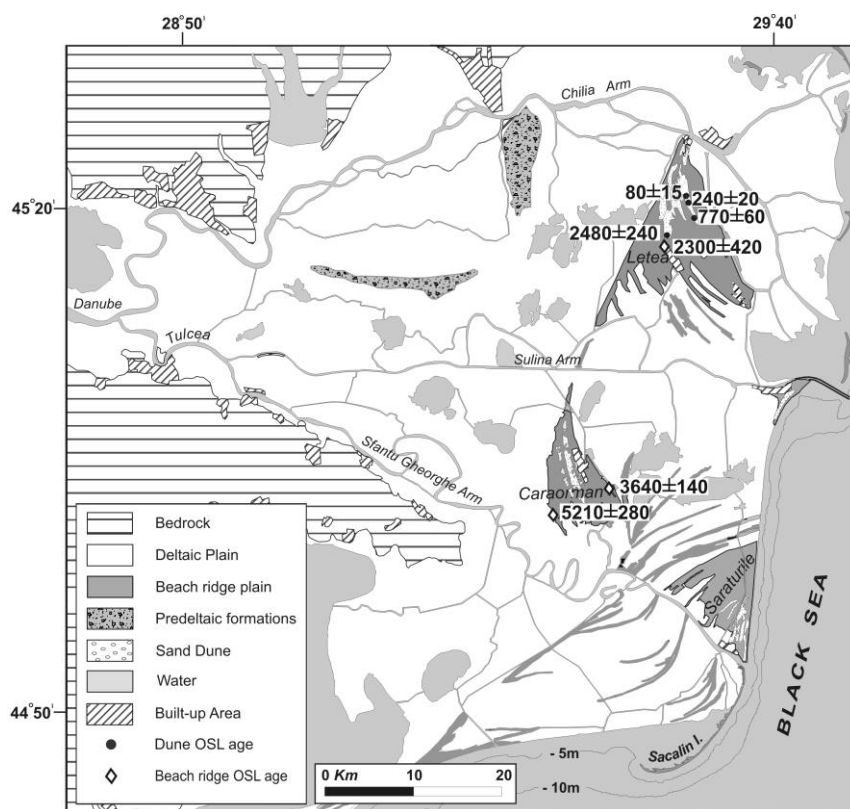


Fig. 1. Danube Delta map and location of OSL samples

(OSL ages are expressed as years before AD 2006; the beach ridge ages on Caraorman are cited from Giosan et al., 2006 and are expressed as years before AD 2004) (after Preoteasa et al., 2009)

According to the most recent studies (Giosan et al., 2006), the Caraorman prograded barrier started (Fig. 1) to form 5600 years ago in association with

the Sfântu Gheorghe (Southern branch of the Danube) arm mouth. Since then, the barrier evolved through deposition of the sediment transported by

the longshore currents updrift of the arm mouth. This pattern of progradation was interrupted when the Sulina arm (the central branch of the Danube) discharge rose and the arm mouth pervaded seaward, delivering Danubian sediments for downdrift deposition. The transition moment from marine to danubian sediment deposition was dated

3640 yrs. B.P. (Giosan et al., 2006). This is also considered the moment when the Letea prograding barrier (Fig. 1) started to form by sediment accumulation updrift of Sulina arm mouth. The preliminary geochronological results yielded from this study confirm this hypothesis.

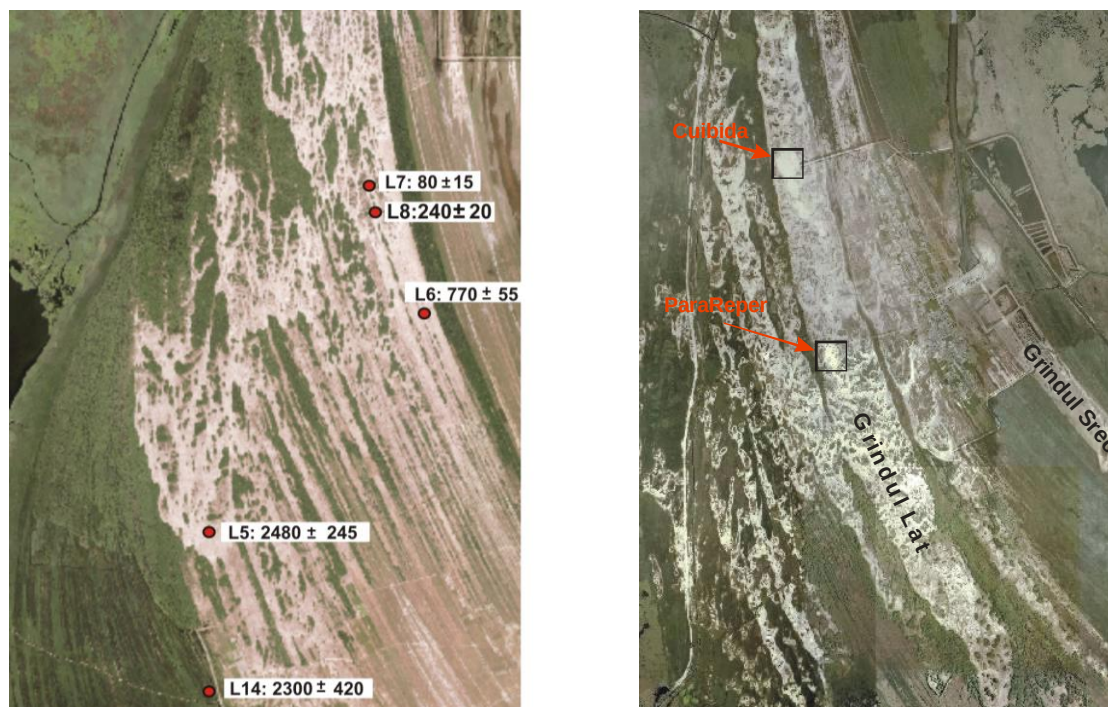


Fig. 2. – a) Orthophotograms (2005) showing the sand dunes on the Letea prograded barrier and OSL ages location (the ages are expressed as years before 2006) and b) the sand dunes from the Caraorman prograded barrier (2006)

A sand sample taken from marine sediments of the beach ridge situated at about 2 km seaward (E) from the Western edge of the Letea prograded barrier, over which the first aeolian deposits overlay, was dated 2300 ± 420 yr B.P. The progradation occurred at different rates, resulting in distinct morphologic patterns: the rapid progradation rates resulted in beach ridge morphology while the slower progradation rate enabled dune ridges building. The mineralogical analysis shows that the aeolian landforms primarily consist in allochthonous clastic sediments originating from the Northern part of the Black Sea Basin, transported by the longshore currents (Panin, 1989; Giosan et al., 2006). Quartz minerals prevail within the marine sand while mica is characteristic for Danubian discharged sediment. The mean grain size of the marine sand varies between 0.22-0.26 mm while the danubian sand is generally finer (< 0.20 mm) (Preoteasa, 2008).

Morphology and chronology of the sand dunes from Letea and Caraorman prograded barriers

Although the two dunefields are only 20 km apart (on an N-S orientated axis) (Fig. 1), they display different patterns of spatial organization of the aeolian features (Fig. 2a,b).

On the Caraorman prograded barrier, aeolian landforms are concentrated within parallel elongated strips, N-SSE oriented, which generally follow the alignment of the underlying marine ridges and of primary foredunes. The dunes are concentrated mostly in the central part.

Commonly the aeolian landforms are represented by dune complexes or individual dunes. Dune complexes regularly consist of parabolic dunes, blowouts and remnant knobs. Most of these evolve under a negative sand budget as no external sediment source to the Caraorman barrier is available, but the reworking of the existent sand. The aeolian features are superimposed on parallel

or sub-parallel ridges, which represent ancient shorelines and which are generally N-SSE orientated. The preexisting topography of these ancient shorelines influenced the pattern of spatial organization of the aeolian landforms as parallel alignments on the Caraorman complex ridge plain. The dunes are northward orientated, with small disturbances induced by wind deflection and channeling effects generated by the existent topography. The observations and measurements we performed on five representative dune complexes reveal a concordance between the steepest slope orientation and the prevalent winds. Thus the extensively existing steepest slopes are NNW exposed, while the resultant drifts direction is $178,1^{\circ}$.

On the Letea prograded barrier, the aeolian features are concentrated within three large (between 5-10 km² each) parabolic-shaped complexes which are situated on the NW part of the barrier, generally N-S orientated (Fig. 2a). The relationship between aeolian landforms orientation and that of the underlying beach ridges is discordant within the southernmost complex and concordant within the central and the northern complexes. Field based observations of the modern morphology revealed the presence of inactive parabolic dunes and nebkhas within the southernmost complex, while the central and northernmost complex display a wide range of overlapping sets of parabolic dunes, nebkhas, blowouts varying from individual parabolic dunes to compounded (nested) parabolic dunes and elongated parabolic dunes. Within the northernmost dunefield the elongated dunes prevail. Previous studies related the formation of elongated dunes with the prevalence of high wind speeds (Gaylord and Downson, 1987; Pye, 1983). Our observations revealed the occurrence of such landforms at locations where the resultant drift direction is concordant with the orientation of the

subjacent topography. The morphology and morphodynamics of each dune complex is probably related to variations of aeolian activity and sediment availability. In contrast, dunes didn't form in the eastern part of the Letea and Caraorman prograded barriers, where numerous beach ridges rapidly formed out of fluvial sediments.

Field based investigations of the modern aeolian morphology revealed the succession of different aeolian activity phases in shaping the modern aeolian landforms. One phase of aeolian modelling of the landforms (Fig. 3) occurred during general low rugosity conditions (bare and grassy areas) and a second one when the aeolian landforms development was constrained by the presence of woody vegetation (*Quercus sp.*) (Fig. 4). Field evidences account for two generations of forest. The oldest trees, were found within the swale areas (Fig. 4). The younger forest generation, whose age was determined as 280 yrs., developed within the interdune areas and blowouts. The highest altitudes of the Letea prograded barrier corresponds to the accretion fronts developed in association with the forest (Fig. 4) (Preoteasa et al., 2009).

Interpretation of the orthorectified aerial photographs of the Letea prograded barrier revealed the discordant superposition of the aeolian landforms over the marine wave-laid sediments as beach ridges. The high discrepancy between the aeolian features orientation and the beach ridges over which they transgress on the southwestern part of the dunefield may account for a long period of aeolian sediment remobilisation. The relation between the resultant migration direction of the dunes (RMDD) and the underneath beach ridge orientation is an important controlling factor of the aeolian morphology development. Elongated parabolic dunes occur where the RMDD is concordant with the subjacent beach ridge alignment.



Fig. 3. Aeolian features developed during low rugosity conditions (bare surface)



Fig. 4. Aeolian accretion front developed in association with the forest

Table 1. Equivalent dose (D_e), dose-rates, and optically stimulated luminescence (OSL) ages of sand dunes (samples Aber111/L5-L8) and beach ridges (sample Aber111/L14) from the Letea prograded barrier (after Preoteasa et al., 2009)

Sample No.	Depth (cm)	Height above aeolian/marine contact (cm)	Water content (%) [†]	D_e (Gy) [*]	$n^{\ddagger}$	External β dose-rate	External γ dose-rate	Cosmic dose-rate [§]	Total dose-rate (Gy/10 ³ yr)	Age (yr) [¶]
L5	110	30	20 ± 5	1.29 ± 0.12	20	0.216	0.126	0.177	0.52 ± 0.02	2480 ± 240
L6	120	30	20 ± 5	0.50 ± 0.03	24	0.315	0.165	0.177	0.66 ± 0.02	770 ± 60
L7	160	400	5 ± 3	0.05 ± 0.01	20	0.255	0.145	0.170	0.57 ± 0.02	80 ± 15
L8	160	450	5 ± 3	0.13 ± 0.01	24	0.242	0.134	0.170	0.55 ± 0.02	240 ± 20
L14	100	-	20 ± 5	2.01 ± 0.36	22	0.451	0.246	0.176	0.87 ± 0.03	2300 ± 420

[†]. Water content expressed as a percentage of the mass of dry sediment, calculated based on measured field values.

^{*}. The D_e shown is calculated as the weighted mean and standard deviation.

[‡]. 'n' is the number of D_e determinations.

^{||}. Dose-rate values (Gy/10³ yr) were calculated using the conversion factors of ADAMIEC and AITKEN (1998) and are shown rounded to 3 decimal places, although the total dose-rates and ages were calculated using values prior to rounding. Central values are given for dose-rates – errors are incorporated into that given for the total dose-rate.

[§]. The cosmic ray dose-rate (Gy/10³ yr) was estimated for each sample as a function of depth, altitude, and geomagnetic latitude (PRESCOTT and HUTTON, 1994).

[¶]. Luminescence ages are expressed as years before 2006 AD, and rounded to the nearest 10 years for ages >100 years, and to the nearest 5 years for ages < 100 years.

Our ages (Table 1), together with the ages produced by Giosan et al. (2006) advocate for a new hypothesis regarding the Letea prograded barrier development according to which at least three phases with different rates of progradation occurred: (1) rapid rates at the beginning: since 3640 ± 140 yr B.P (Giosan et al., 2006) to 2300 ± 480 yr B.P., (2) slow rates since 2300 ± 480 to 900-1200 yr B.P (uncertain radiocarbon age: Giosan et al., 2006), (3) rapid rates since 900-1200 yr. B.P. until the initiation of the Chilia secondary delta (Fig. 1). The different rates of progradation resulted in distinct morphological features: *beach ridges*, 6 km wide, developed as a result of high prograding rates and *foredune ridges*, emplaced during the slow prograding phase. The different widths of foredune ridges area in the northern (cca. 1 km), central (cca.

2 km) and southern (4 km) parts suggest different mean progradation rates varying from less than 1m/yr on the northern part to 2-3 m/yr on the southern part.

The concentration of the largest aeolian sand volumes on the N, NW part of the barrier could be interpreted as the alongshore variation of the foredune volume in relation with the LST (which is the main sediment source for Caraorman and Letea dunes). This alongshore foredune volume variability hypothesis is sustained by the reiteration of the same pattern over the entire width of the barrier (occupied with dunes) which developed within a time interval of 1600-1800 years. Their localization in the N, NW part of the barrier could have been promoted by the position in relation with the alongshore currents sediment transport capacity and shoreline exposure

to the prevalent winds and waves. Subsequently, the rugged aeolian morphology concentrated on the northern part of the barrier was easily reworked by

the prevalent northern winds and redistributed southward (downdrift) to finally form just a thin veneer of aeolian sand overlying the beach ridges.

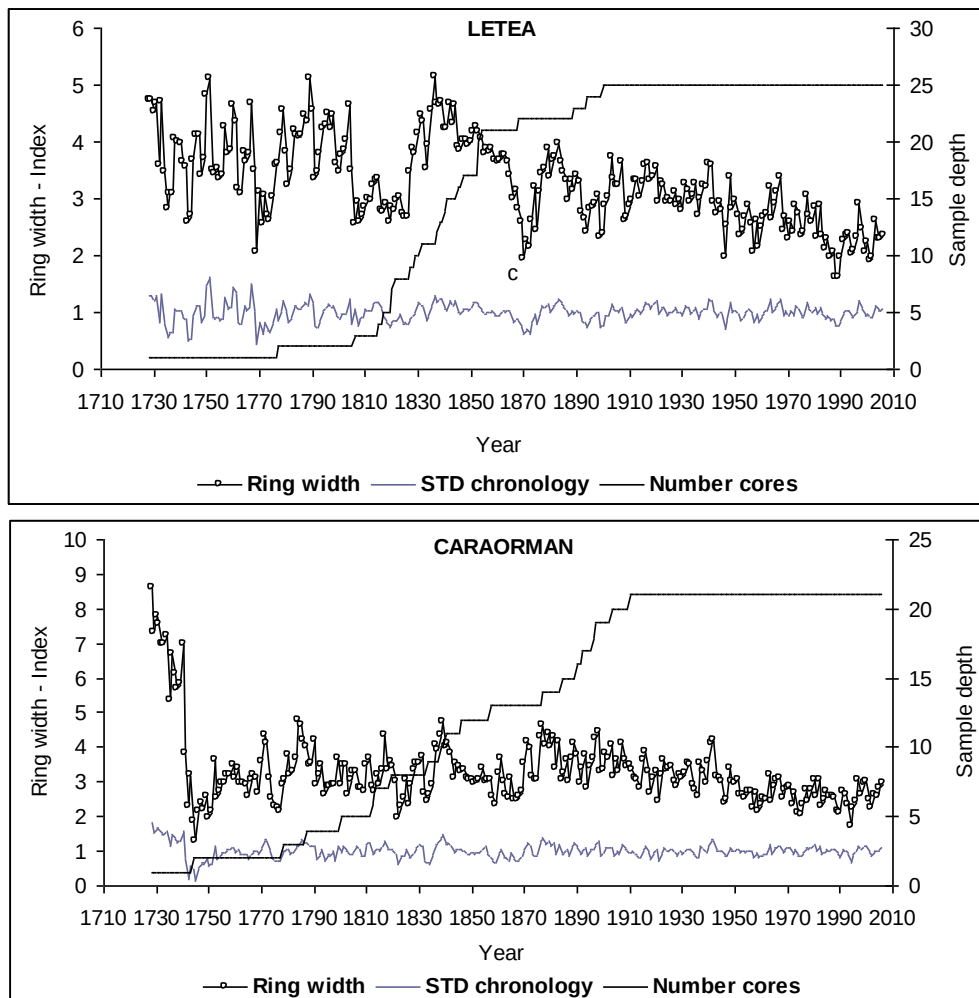


Fig. 5. Dendrochronological analysis of Letea and Caraorman forest

The stable behaviour of the shoreline on the northern sector permitted an important sedimentary transfer from nearshore to beach-dune system enabling thus the development of the large foredunes. The altitude of beach ridges is almost similar over the entire prograded barrier suggesting the fact that a steady wave climate and small sea level changes occurred during the Upper Holocene (Giosan et al., 2006).

The integrated geochronologic and morphometric analysis of some representative sand dunes from the Letea and Caraorman prograded barriers revealed a high rate of sand transport averaging 0.68 m/yr within the last 240 yrs.

The tree ring growth analysis presents more time intervals when the tree development was unusually slow (Fig. 5). Some of these intervals are correlated to drought intervals as suggested by tree

ring growth characteristics and as recorded by hydrographical and meteorological measurement (e.g. 1870, 1946-1947), while some others do not correspond to drought records. The slow tree ring growth intervals that the dendrochronological analysis highlights are: 1767-1776 (Letea) and 1770-1778 (Caraorman); 1802-1828 (Letea) and 1806-1814/ 1820-1828 (Caraorman) and 1982-1995 (Letea) and 1982-1995 (Caraorman) (Fig. 6). Significant are the duration of these intervals (9-20 years) and the synchronous biological processes that occur within the forests on the two prograded barriers.

Letea 8 age fits well within the interval of slow three development recorded between 1767-1776 suggesting the high sand transport rates as a potential explanation.

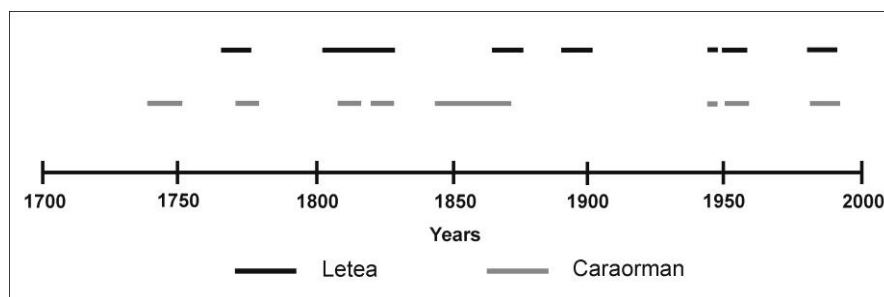


Fig. 6. Slow tree ring development time intervals in Letea and Caraorman forests

Conclusions

Preliminary OSL ages of Letea dune field indicate their formation during several episodes. They formed as successive seaward advancing foredune ridges between 2400 yrs and 770 yrs ago. The first aeolian landforms were synchronous with the process of eastward barrier progradation. Their evolution involved a primary phase when they developed as a result of the sedimentary exchange within the beach – dune system resulting into parallel foredune ridges. The constant barrier progradation lead to sedimentary flux decoupling between foredune ridges and the adjacent beach. A second phase followed consisting in several episodes of aeolian reworking of sediments already existing into the system. The interplay between the sedimentary flux and the woody vegetation resulted into the modern configuration of the aeolian landforms on which several episodes of aeolian activity could be detected. The most recent aeolian activity episodes occurred 240-70 yrs ago possibly controlled by the climatic variability during the Little Ice Age. The tree ring growth analysis presents more time intervals when the tree development was unusually slow. Some of these intervals are correlated to drought intervals as suggested by tree ring growth characteristics and as recorded by hydrographical and meteorological measurements (e.g. 1870, 1946-1947), while some others do not correspond to drought records. Letea 8 age fits well within the interval of slow tree

development recorded between 1767-1776 suggesting the high sand transport rates as a potential explanation.

The correlation between NAO index, precipitations and tree ring development revealed slow tree ring growth during the positive phases and rapid development during the negative phases.

The marine origin of the dune substrate justifies a regressive (prograding) barrier development model. Other papers dealing with aeolian landscape reconstruction elsewhere in Europe, report intensive episodes of aeolian activity within the last 250 yrs. For example, Clarke and Rendell (2006) documented a recent dune building episode in Portugal, occurring between 1770-1905 AD, Murray and Clemmensen (2001) identified the most recent aeolian activity phase at Thy, Denmark, starting earlier than 200 yrs ago, and Clemmensen et al. (2007) established the chronology of the last aeolian activity phase between 1640 and 1900 AD.

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Rockfall Hazard Assessment

Case study: Lotru Valley and Olt Gorge*

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Key words: rockfalls, hazard assessment, roads, heavy rainfall

Abstract. Rockfalls - a natural hazard – can occur isolated in our country but they produce serious damage, hurting and even killing peoples. Despite all financial losses, rockfalls were not studied in Romania by the geomorphologists and engineering geologists. Those events are strictly linked with those climatic one and also with human activities which created slope instability. In this paper we tried to assess this kind of relations and also magnitude – cumulative frequency relationships and return periods. We tried also to estimate the probability of rock falling onto a moving vehicle using one simple formula made by the Australian Geomechanics Society. The two major events, the damages and people injuries caused by rockfalls in the last 8 years (after 2001) are described in detail.

Introduction

Rockfalls are present-day geomorphic processes spread in all mountain areas (Rapp, 1960; Whalley, 1984). They may occur both in natural (untouched slopes by human activities) and cut slopes (where a linear infrastructure exists). In this paper we will try to assess only the cut slopes along two major transport corridors.

Rockfall is defined as „a fragment of rock detached by sliding, toppling or falling from a vertical or subvertical cliff, before proceeding downslope by bouncing and flying along parabolic trajectories or by rolling on talus or debris slopes” (Varnes, 1978) or „the displacement of a single fragment or several pieces” (Evans & Hungr, 1993).

In Romania studies on rockfalls were not done neither by geomorphologists nor by engineering geologists as there were on landslides (soil slips, mudflows and snow avalanches), perhaps because the first occur on small and isolated areas.

When rockfalls may produce victims or even cause damage, destroying roads, railways, electricity lines, water and pipelines etc. they are considered natural hazards. Although on the rockfall hazard many kind of rockfall hazard assessments along transportation corridors can be found in the foreign literature (Ritchie, 1963; Pierson et al., 1990; Pierson & Van Vickle, 1993;

Baillifard et al., 2003; Hantz et al., 2003; Budetta, 2004; Guzzetti et al., 2004; Uribe-Etxebarria et al., 2005), the assessment is still very difficult to achieve.

In most cases the greatest problem is represented by the lack of any data because only the most important and destructive rockfall events are noted in the institution files.

Study area

For analysis we selected two important transportation corridors affected every year by many rockfalls. The first site is the Lotru Valley (the sector between the Voineasa and Brezoi localities) crossed by the national road 7A and the second one is the Olt Gorge crossed by the European road E 81 (Fig. 1), which is also the road the most affected by this kind of geomorphic hazard from the whole country. We selected these two corridors because the rockfall frequency and also the traffic intensity are high. The length of the corridors is almost 100 km (39 km has the first and 50 km the second).

From an administrative point of view the two corridors are part of the Vâlcea County and only a few kilometers from the northern sector of the Olt Gorge is part of the Sibiu County.

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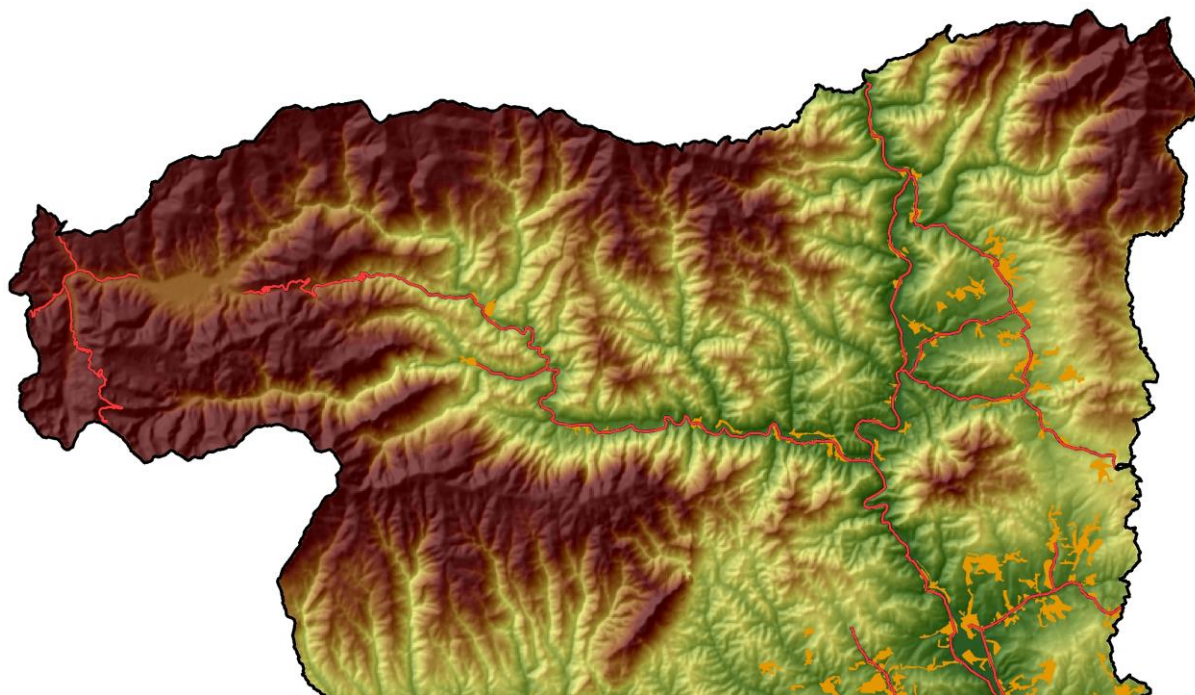


Fig. 1. The position of the two main transport corridors.
The W – E axis is the National Road 7A and the N – S red line represents the European Road 81

Geological settings

The geology is very important when we try to assess any hazard, because it influences the magnitude of events through the complexity of lithology and the presence of joint sets.

Both sectors are cut into metamorphic rocks (schists, augen gneiss, paragneiss) and less into sedimentary rocks (the sector between the Valea lui Stan and Brezoi localities as well as between the Căineni and Golotreni localities), which are made of conglomerates, breccias and sandstones and are part of the Brezoi – Titești sedimentary basin (Savu & Schuster, 1975; Savu et al., 1977; Popescu et al., 1977; Ștefănescu et al., 1982). The metamorphic rocks are included in the Getic Nappe and the Supragetic Nappe.

Although the hardness of these rocks is higher, the joints set (faults, foliation, bedding plane) have a major role in the detachment of rock blocks or fragments (Hencher, 1987). The most important fault of the entire region is the Brezoi Fault which has 80 km in length and an uplift of about 1000 m along the discontinuity plane (Ghika – Budești, 1958). It can be observed from the northern slope of the Cozia Massif and then at the bottom of Căpățâni Mountains up to the Malaia Valley.

The joint sets are very important when researchers want to assess rockfalls to a local scale

(microslope level for example). Generally large rockfalls are triggered under planar, wedge and overhang failures (Fig. 2). When joints have random orientation, the rock mass strength is lower and rockfalls occur.

Rock weathering is another geological factor which contributes at the failure mechanism. Although most part of the region is made of metamorphic rocks, the failure mechanism is possible due to physical and chemical weathering and also to the differential erosion. Many slopes are affected by these three processes especially in the sector between the Brădișor Dam and Voineasa locality (Lotru Valley) and between the Brezoi and Cozia localities (Olt Gorge).

Geomorphological settings

Both valleys are limited by steep slopes. In the cross sections along these valleys, slopes can be separated into two sections: the bottom part which is generally cut for road and railway bed, very steep (often vertically) without natural vegetation or with sparse vegetation and the upper part which is not affected by human activities but generally also very steep. Usually the cut slopes are 10-30 m high but the natural slopes can reach more than 100-150 m in height. The slopes aspects are N and S for the Lotru Valley and E and W for the Olt Gorge.



Fig. 2. Types of failure mechanism: overhanging block in sandstone (left) and topple in paragneiss

Climate settings

Hoek (2007) concludes that „rockfalls are generally initiated by some climatic or biological event that causes a change in the forces acting on a rock”. Here can be included: „pore pressure increases due to rainfall infiltration, erosion of surrounding material during heavy rain storms, freeze-thaw processes in cold climates, chemical degradation or weathering of the rock, root growth or leverage by roots moving in high winds”.

As in the case of the landslides from our study area (debris flow), there are few examples of rockfalls generated by heavy rainfall. Firstly, we can note many events which occurred during the August 10-11' 2007, then October 22-24, 2007 and March 08, 2009. The rockfall deposits were observed between the Brădișor Dam and the Voineasa Spa on the carriage way. In these regions, rockfalls occurred under heavy daily rainfall usually higher than 35-40 mm (Fig. 3).

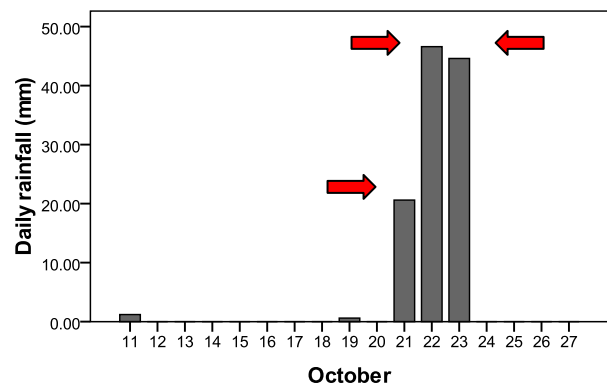
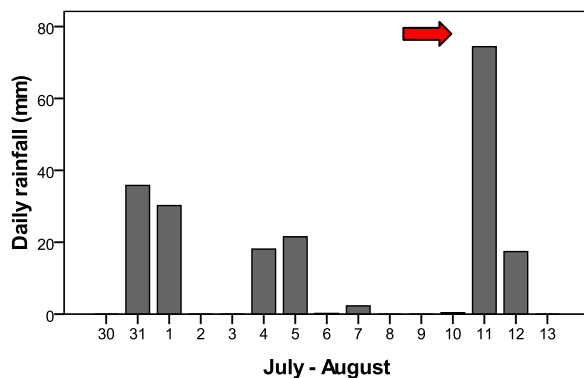


Fig. 3. The daily rainfall for two periods with multiple rockfall occurrences. Red arrows indicate days with rockfalls

Another example includes also the antecedent moisture from regolith. A big debris/rock fall was triggered in March 26, 2006 and destroyed an old motel in the Gura Latorîței point. The moisture from the regolith triggered an initial slide and then a huge debris/rock avalanche (Fig. 7).

For the period 2003 - 2007 we found a good relation between monthly rainfall and rockfall events (Fig. 4). The relation is not valid for all months with high rainfall; few explanations are credible here:: (1) the deficiency of inventory (it is

possible that rockfalls occurred in the months with the highest monthly rainfall but they were not registered in the official files); (2) large rockfalls did not occur, because before the month with a high monthly rainfall amount, only small events (fragments by fragments) occurred; (3) rainfall was distributed uniformly during the month that induced a low daily rainfall and made data difficult to analyze. It is known in practice that rockfalls are very often caused by rainfall and their relationship is weak (Corominas, 2000).

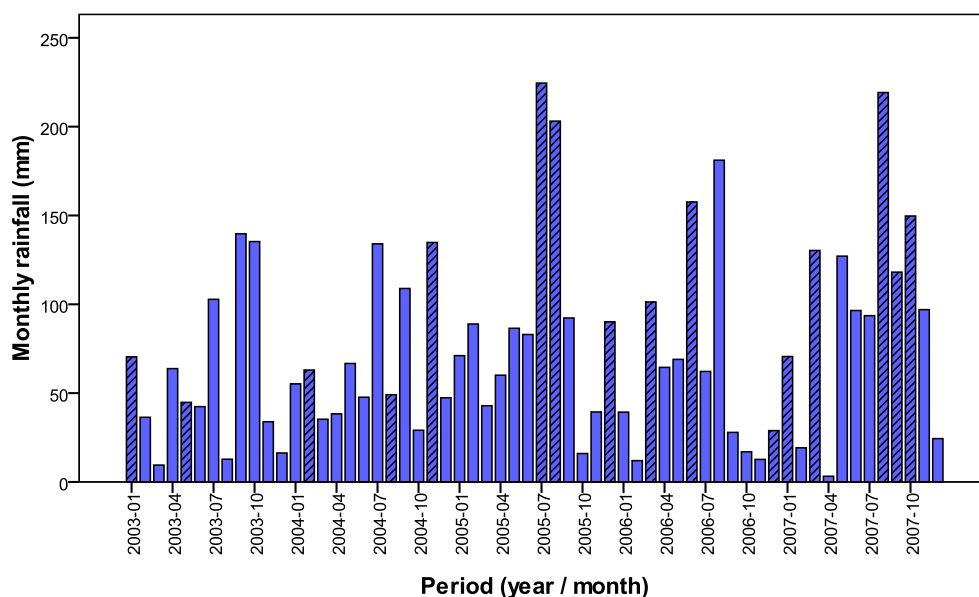


Fig. 4. Monthly rainfall and rockfall events between 2003 and 2007. The dashed columns refer to months with rockfalls

Regarding the rockfall distribution over a year (monthly frequency) it is clear that we have high frequencies in two periods. Firstly, during the months which had high daily rainfall in the last years (after 2003) especially July and August (Fig. 5). Although the rainiest months are May and June, our data base does not indicate a high frequency for these months, perhaps because of data lacking and the short period of inventory.

Secondly, it is the case of the transition seasons with freeze-thaw cycles, especially November, February and most of all March (Fig. 5). The last month has a high frequency of rockfalls due to the events that occurred on March 8, 2009, when we found 17 points with blocks on the road (Fig. 8). In our data base there are many multiple-occurrence events in both valleys.

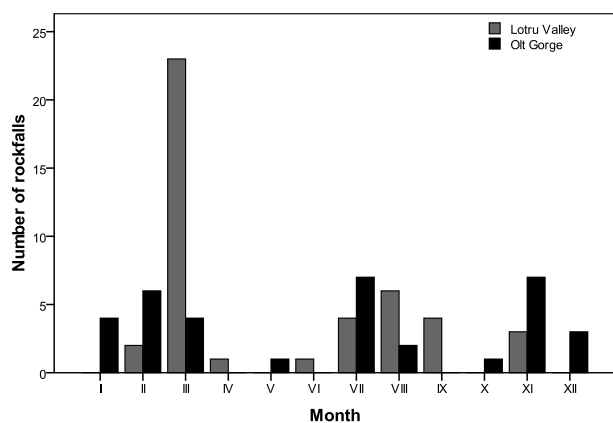


Fig. 5. Monthly rockfall frequency in both valleys

Hazard assessment – methods and data base

Because of the lacking information, we tried to create our own data base. In order for this to be more complete, we used the data from a few sources: the National Roads Administration from Râmnicu Vâlcea, online newspapers and our field observation and measurements. The period covered only 6 years (2003-2009) for the Olt Gorge and 5 years for the Lotru Valley (2004-2009). The total number of rockfall events is 79 (only 63 of them with registered volume) but the real number is much higher than this (it is very possible rockfall events were more than 300). The rockfalls volume ranges between 0.1 to 10.000 m³ with a mean of about 415 and a standard deviation of 1558.

Magnitude and frequency

In rockfall studies, like in others natural hazard, one of the most important aspects is magnitude – cumulative frequency relationships (Hungry et al., 2008). Magnitude is defined here as the volume of rock and debris displaced by a landslide (Hungry, 2005; Hungry et al., 2008) and the frequency can be expressed as the number of events of a given magnitude in a given unit of time (Alexander, 1993).

Magnitude – cumulative frequency relationships were adopted from seismology (Gutenberg and Richter, 1949) and applied in rockfalls and debris flows assessment by geomorphologists and engineering geologists (Wieczorek et al., 1995;

Hungr et al., 1999, 2008; Dussauge-Peisser et al., 2002; Chau et al., 2003).

We applied this relationship for events which occurred in both valleys – Lotru and Olt – which are also two transport corridors. The annual frequency for a rockfall of a certain magnitude can

s

be read from the graphic curve (Hungr et al., 2008). As can be seen in the graphs (Fig. 6 a, c) the curves in both cases are similar and the low magnitudes have a high frequency and, vice versa, high magnitudes have low frequency.

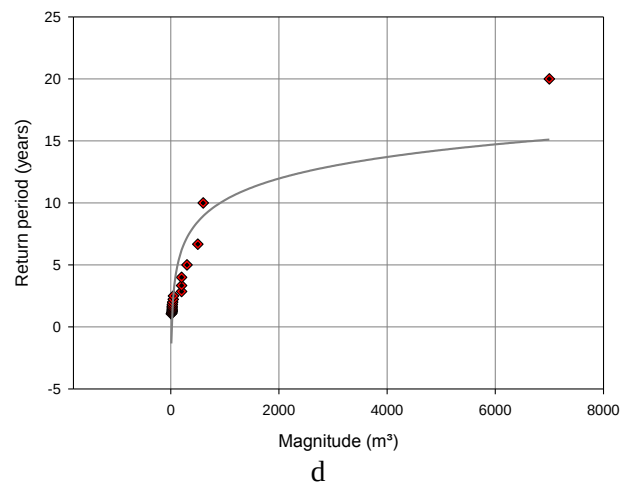
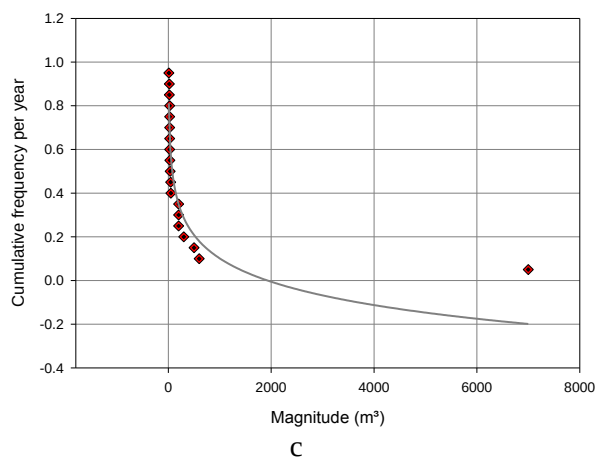
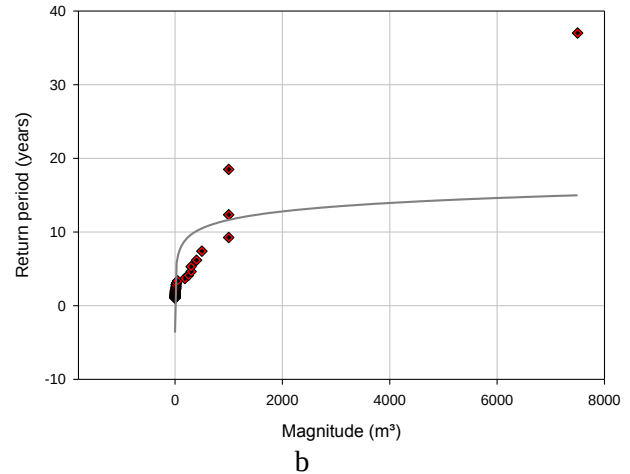
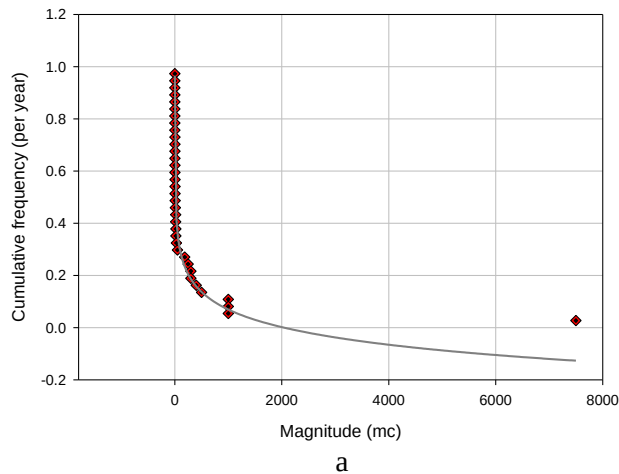


Fig. 6. **Magnitude – cumulative frequency relationships and return period for the Lotru Valley (a, b) and the Olt Gorge (c, d)**

After obtaining these relations, it is very easy to estimate the return period, as can be seen from Fig. 6 b and d. Because the high frequency events are insignificant and usually they do not damage the infrastructure we are interested especially in large events return periods because, generally, the latter events destroy roads, railways and buildings. The return period for the event with a high magnitude ($>7000 \text{ m}^3$) is higher and is approximately 15 years (Fig. 6 b, d).

But this data inventory is not complete and therefore the graphs above may be regarded with skepticism. Anyway, the magnitude – frequency relationships remain an important tool when geomorphologists and engineering geologists try to

assess the hazard of rockfalls, especially along linear infrastructure.

Câineni and Gura Latoriței – two major events in the last years

In the last years many events occurred which produced serious damage to roads and railways and also hurt a few people inside the cars that crossed the valleys. Two of these were large rockfalls that detached thousand of square meters of rock downslope.

One of these events occurred in December 28, 2005 in the Northern sector of the Olt Gorge, along

the European Road 81. Here, more than 7000 m³ (10000 m³ from another source) were detached from a 40-50 m high cliff and covered the railway and the road on a more than 30 m distance (Fig. 7 a). This was a typical rock avalanche. The European Road was turned-off between December 28, 2005 and 11 January, 2006 and the financial loss was also very high. At the time when the rock avalanche was triggered, a train was in the proximity of this point.

After this event the Ministry of Transport invested a lot of money for rockfalls remedial measures and also for continuing with extending the existent railway tunnel at both ends. After many years of investments the people still work in this sector and the financial loss scaled up and exceeded millions of Euros.

In the case of the European Road it was very easy, because in this sector the valley is relatively

large and the workers could deviate the river towards its left bank by 4-5 m. After that, they constructed from beginning a new road bed on the right bank which is a few meters away from the bottom of the slope.

Another large event occurred in the Lotru Valley along the National Road 7A in March 26, 2006 at the Gura Latoriței point. Here, the debris/rock avalanche included more than 7500 m³ (Fig. 7 b). The material included rock blocks (generally smaller than 50 cm in diameter), soil, deluvial deposits and many trees, and covered two roads: the National Road 7A and the Local Road 175 which stopped the traffic. An old motel was also completely destroyed and a few people nearly died because at that moment they were inside.



a



b

Fig. 7. Two major landslides: (a) the Căineni rock avalanche (Photo: Mediafax) and (b) the Latorița debris/rock avalanche

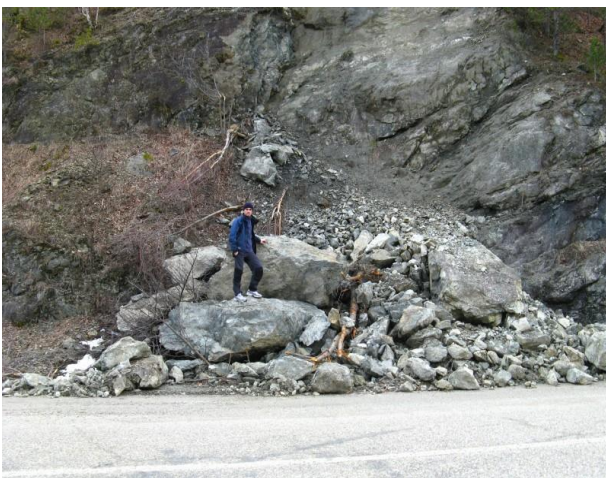


Fig. 8. The rockfall event, from March 8, 2009, along the National Road 7A (Lotru Valley). The blocks in the left photo have about 20 m³ and covered part of the road (the photo was taken after the authorities unblocked the road); the boulder from the right photo has a 1.5 m diameter

Probability of rock falling onto a moving vehicle

To calculate the probability of rock falling onto a moving vehicle we adopted the formula from the Australian Geomechanics Society (2000):

$$P(s) = 1 - \left(1 - \left(P_{(S:H)}\right)\right)^{N_R} \quad 1)$$

where:

$P(s)$ = probability of one or more vehicles being hit;
 $P_{(S:H)}$ = probability of a vehicle occupying the portion of the road onto rock falls;

N_R = number of rockfalls/day,

and

$$P_{(S:H)} = \frac{N_v}{24} \cdot \frac{L}{1000} / V_v \quad 2)$$

where:

N_v = number of vehicles/day;

L = length of the vehicles (m);

V_v = velocity of vehicles/hour (km/hour).

We selected the National Road 7A to apply this equation. As can be seen, the results depend on the number of vehicles per day, the length of vehicles, the velocity of vehicles and the number of rockfalls per day. For this reason we applied the formula for a velocity between 50 and 100 km/h and a length of cars of 5, 7.5, 10 and 12.5 m. The daily traffic is about 1350 vehicles/days and the number of rockfalls is 1/month. The daily traffic information was obtained from the National Road Administration in Râmnicu Vâlcea which has installed a monitoring video camera in the Malaia village and the number of rockfalls was estimated from the data base inventory.

From the graphics it can be observed how the probability of one or more vehicles being hit and the probability of a vehicle occupying the portion of the road onto rockfalls decrease proportionally with the decreasing of velocity. For a velocity of 50 km/h $P(S:H)$ is 0.0056 and $P(s)$ is 0.000079 and for 100 km/h $P(S:H)$ is 0.0028 and $P(s)$ is 0.000039 (Fig. 9). For this estimation we assumed that the daily traffic is constant and the distance between vehicles remains equal and no changes appear in this time (this is impossible in practice). Therefore this estimation is partial.

Roads are very frequently damaged due to rockfall events. People or cars are rarely hit by blocks when they cross the corridors transport. In the last years (after 2000) we can note two special events which seriously hurt people. The first event occurred in the second sector of the Olt Gorge

(between the Brezoi and Cozia localities) and hit a police agent who drove onto Râmnicu Vâlcea. One block fall onto his car and hurt him very seriously. After that event the man's recovery was difficult and he remained with some physical lesions (and of course psychic damage) for the whole life. In 2002 a car with 3 people on board was also hit by a rockfall event, 2 of them being seriously hurt and another very easily. In 2008 two cars were hit very easily and nobody was hurt. From our information nobody died due to rockfalls events.

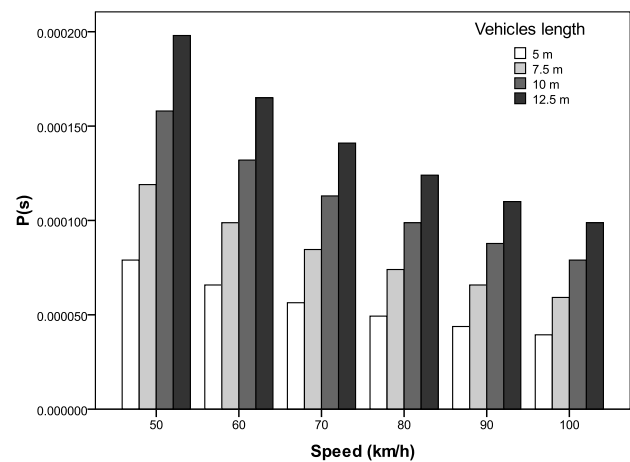


Fig. 9. The probability of one or more vehicles being hit $P(s)$ taken in account different vehicle lengths and different speeds (National Road 7A)

Damage, hurt and killed people due to rockfalls

Rockfalls destroy very frequently the roads and railway and produce huge financial losses. Even when they occur without destroying the roads, they still produce financial losses because they may interrupt periodically the traffic and money are spent to release the way.

Conclusions

Quantitative estimations for rockfalls which occur in transport corridors are very important. One of the most important aspects is the lack of a data inventory because it is impossible to make the hazard map without this step. A problem is the heavy access to authority archives. If in the future those institutions which are involved in the administration of national and European roads will create a landslide data base inventory, the geomorphologists and engineering geologists will have a start point when trying to assess the landslide hazard.

Most of the rockfall events occurred in two distinct periods: in summer, when heavy daily rainfall occurred and at beginning and ending of

winter as a consequence of freeze-thaw cycles, therefore more accurate meteorological data are needed.

Acknowledgments

The studies were financed by CNCSIS, grant TD no. 321/2007. I would like to thank the National Road Administration of Râmnicu Vâlcea and to Ms.

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MISCELLANEA

13th Belgium – France – Italy – Romania Geomorphological Meeting *Landscape evolution & geoarchaeology* Porto Heli (Greece), June 18-21, 2008

The 13th meeting was held in Porto Heli, Greece, which was a proper chosen location (considering the specific task of the symposium) by the organizers – French and Greek geomorphologists. One of the aims of this meeting of the geomorphologists from the four countries on the Greek territory was to co-opt the Greek colleagues inside the “alliance”. The 14th Italy – Romania – Belgium – France – Greece meeting will take place in Romania in 2010.

Presentations (40 oral presentations and 25 posters) of 19 and 20 June were thematically grouped as follows:

- Long term Paleo-environmental dynamics and archaeo-logical settlements (chairing: E. Fouache - K. Pavlopoulos; M. de Dapper-N. Evelpidou; M. Dermitzakis-N. Evelpidoy);
- Coastal Environments (chairing: J. Ph. Goiran – G. Livadidis; S. Poulos-G. Papatheodorou);
- Geoarchaeological practices and methods (chairing: V. Kapsimalis – L. Lespez; K. Pavlopoulos – A. Vott);

- Human activities and landscape (chairing: F. Vermeulen-N. Josan; F. Dramis-L. Stamatopoulos; J.-L. Pena-Monne – A. Livaditi);
- Archaeoseismology (chairing: St. Stiros-I. Mariolakos);
- Fluvial and spoles dynamics (chairing: F. Grecu-A.Vott; K. Vouvalidis- D. Sakellariou).

The one day field trip (21 June) revealed the most specific features regarding the geomorphology, geology, civilization and culture of Argos Plain and Pelopones around Argos Bay, such as: ancient Tiryns and the Mycenaean dam; channel of Megalo Rema; the geoarchaeology of the area of the Acropolis of Mycenae, tomb of Agamemnon; Acropolis of Midea; Nafplio; Argos; Hadrian’s aqueduct; aluvial fan of Xerias in the Argos.

High academic statute of the presentations and field trip has proven that changing ideas is necessary and useful for a direct cognition of the land, culture and civilization of the peoples.



IAG/AIG Regional Conference on Geomorphology
Landslides, Floods and Global Environmental Change in Mountain Regions
 Braşov, September 15-25, 2008

The IAG/AIG Regional Conference on Geomorphology having as theme ***Landslides, Floods and Global Environmental Change in Mountain Regions*** took place during September 15-25, 2008, at the Faculty of Silviculture and Forest Engineering of the Transylvania University of Braşov. The meeting was held under the authority of the International Association of Geomorphologists, the Romanian Association of Geomorphologists (AGR) and the Carpatho-Balkan Geomorphological Commission (CBGC), was locally organized by the Institute of Geography of the Romanian Academy under the leadership of Prof. Dan Bălţeanu.

The Conference included plenary lectures of some of the best world's geomorphologists, 12 scientific sessions of oral and poster papers, including special Workshops and IAG Working Group sessions, pre- and post-conference field trips and an International Summer School on Natural Hazards for young geomorphologists. In addition, special activities took place, like the meetings of the IAG Executive Committee, the Romanian Association of Geomorphologists and the Carpatho-Balkan Geomorphological Commission, IAG Training Activities and cultural visits for accompanying persons. The two special workshops as well as the International Summer School were financially supported by the Romanian Ministry for Education, Research and Innovation.

At the Opening Ceremony, speeches were held by Prof. Andrew Goudie, President of the IAG/AIG, Prof. Dan Bălţeanu, director of the Institute of Geography of the Romanian Academy and Prof. Ioan Vasile Abrudan, dean of the hosting Faculty. Prof. Mauro Soldati, treasurer of the IAG, offered three grants of 500 Euros each to the winners among young geomorphologists applying for participating in this Conference.

The plenary lectures were presented by Prof. Andrew Goudie (Oxford University Centre for Environment, UK), Prof. Dan Bălţeanu (Institute of Geography, Romanian Academy, Romania), Prof. Monique Fort (Université Paris, France), Prof. Michael Crozier (Institute of Geography, School of

Earth Sciences, New Zealand), Prof. Irasema Alcántara-Ayala (UNAM, Instituto de Geografía, Mexico), Prof. Franck Audemard (FUNVISIS, Earth Sciences Dept., Venezuela), Prof. Jean-Noel Salomon (L'Université Michel de Montaigne-Bordeaux, Institut de Géographie, France), Prof. Vladimir Boynagryan (Yerevan State University, Armenia).

Over 150 specialists, mainly geomorphologists, from 30 countries, attended the international conference and 111 abstracts were accepted. The scientific sessions included six general sessions, two special Workshops (Exploratory Workshop on *Climate Change and Extreme Events in Mountain Regions* and a Workshop on *Hazard Assessment in Mining Areas using GIS*) and four IAG Working Group sessions (*Geomorphological Hazards, Geomorphology and Society, Human Impact on the Landscape, as well as the Third Joint Council Meeting of the IAG Carpatho-Balkan-Dinaric Regional Working Group and the Carpatho-Balkan Geomorphological Commission*).

The accepted abstracts were published in a special volume of 121 pages. The presentations focused upon the impact of landslides and floods on settlements, infrastructure and environment in the context of the global climatic change; hazard mitigation measures, modern methods for the evaluation of landslide failure mechanism and monitoring of the impact of landslides.

Four field trips took place offering interesting sights into the Romanian geomorphological features: a pre-conference one-day (54 participants), an afternoon (55 participants) and two post-conference field trips: *Vertical Zonation of Floods and Mass Movements in the Carpathian Mountains* (21 participants), *Large Landslides, Floods and Seismic Activity* (19 participants). The field trip guides were published together in a book of 128 pages.

At the **International Summer School for Young Geomorphologists *Environmental Hazards and Sustainable Development in Mountain Regions*** (September 20-25, 2008, Pârlăgele, Romania), ten young PhD and Master

Students from six different countries (Estonia, Malta, Slovenia, Hungary, Turkey and Romania) participated.

The main goals of the Summer School were: identification, inventory and analysis of geomorphic and hydrologic hazards and their effects on the natural and social environment in the context of global climate change.

Prof. Michael Crozier from New Zealand, invited at this edition of the Summer School, as well as Prof. Dan Bălteanu (Romania) and Markus Holub (Austria) held courses on landslides and other geomorphic processes, on floods and engineering structures.

The School participants also enjoyed the presence and the interesting discussions with some other main figures in geomorphology, like Prof. Monique Fort (Université Paris 7 – Denis Diderot,

France), Prof. Franck Audemard (FUNVISIS, Earth Sciences Dept., Venezuela), Prof. Piotr Migon (University of Wrocław, Poland) or Prof. Denes Loczy (Institute of Geography, University of Pécs, Hungary). This was possible, as the first three days of the Summer School were spent having the same scientific programme as the field trip *Large Landslides, Floods and Seismic Activity*. The field trips focused on explaining the causes, the evolution, the specificity and the environmental effects of some landslides, mudflows, river floods and flash-floods, active faults and mud volcanoes.

The young researchers presented their works and discussed their ideas with experts in the field of landslides, applied geomorphology, natural hazards, mountain geomorphology, rock control in geomorphology, active tectonics, land evaluation and land use etc.

Marta Cristina JURCHESCU
Conference Secretary

KAZAKHSTAN Expedition 2009

KAZAHSTAN 2009 Geographical Expedition took place in the south-west of Kazakhstan Republic, in the northern region of Aral Sea, during the month of August 2009. The expedition was organized by DS Spirit Association and sustained by the Faculty of Geography and the Research Center Land Degradation and Geomorphological Dynamics - University of Bucharest, Romanian Geographical Society, Emil Racovita Institute of Speleology, National Geographic Romania Magazine. The expedition grew from the idea of better understanding the realities of Aral Sea region and also of the desert and steppe regions the expedition covered.

From the nine participants in the expedition, three were geographers: PhD Anca Munteanu and Veronica Darmiceanu from Faculty of Geography, University of Bucharest and PhD Igor Sirodoev from Institute of Geography in Chisinau. Biologists, architects, photographers and engineers completed the team. Each of the nine team members tried to observe and interpret according to its academic background, the different aspects encountered during the expedition, gathering a lot of information and also a rich imagistic material.

The main objectives of the expedition were: to gather information, to immortalize and to make known objectives of scientific (geomorphologic, geographical, etc.), socio-cultural and touristic interest from the west of Kazakhstan Republic; to produce a photo-video documentary upon the current situation of Northern Aral Sea; to investigate the evolution of the environmental components, uttering the dramatic drop in sea level and the changes that occurred in the landscape; to investigate the geomorphologic processes and relief characteristics, the changes determined by the

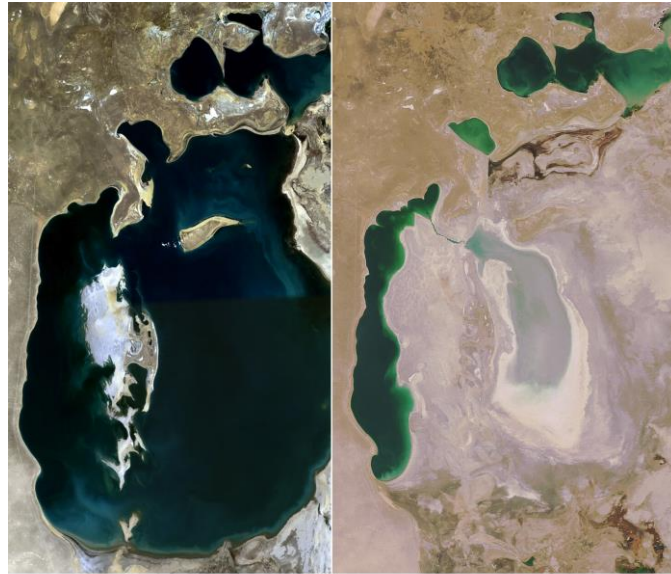
global warming, the expansion and intensity of desertification processes and its effect on the local communities, the evolution of the desert relief ; to analyze of the current situation of the region.

The expedition, benefiting of 4WD cars, covered approximately 9500 km through Romania, Ukraine, Russian Federation and Kazakhstan. Part of the observations were made along the way, in the Volhynian-Podolian and Donetsk Upland, north of Azov Sea, Volga and Ural Delta, Barsuki and Prearal Deserts, Mukhadzhar Hills.

In the North Aral Sea region, several routes in the eastern, north-eastern and southern part of the sea were followed. We crossed the bridge over Syr Darya and the Kokaral dike, which separates the Northern Aral from the Southern Aral Basin, we walked on the Kokaral peninsula and on the former seabed invaded by the desert, we were amazed by the former coastline modeled by the current subaerial processes and by the ships unstuck in the sand, witnessing the disaster.

The disaster seems to loose its strength in the North Aral Sea. The construction of Kok-aral Dike in 2005, in order to prevent the outflow from the North Aral in the South Aral, already gives results. The sea level raised by 12 m and the former port of Aralsk, which at one point was 100 km away from the sea, is now just 25 km from the water.

The whole expedition represented a great opportunity to better understand the reality of desert and steppe environmental conditions, which replaced, because of the uncontrolled anthropic activity, one of the largest lakes on Earth. It was also an opportunity to reach less "touristic" places from the ex-soviet space, and to travel through the wilderness of steppes from the "land of nomads".



July - September, 1989

October 5, 2008

Comparative evolution Lake Aral**The bottom and former cliffs of Lake Aral, invaded by the desert****A stranded ship. In the background one can notice the level Lake Aral had in 1930**

Dr. Anca MUNTEANU

The International Symposium of the Swiss Geomorphological Society, 3-5th September 2009, Olivone, Canton of Ticino, Switzerland

The beginning of autumn was marked in the Swiss geomorphologic' calendar by the Symposium of the Swiss Geomorphological Society, a scientific event with an international participation, having the theme "The Alpine Geomorphology: between patrimony and constraints". The picturesque small town of Olivone situated in the sunny Italian-speaking canton of Ticino, on the Blenio Valley, was the place selected this year for an interesting geomorphologic meeting. About 40 participants joined the symposium, most of them Swiss geomorphologists but there were also specialists from France, Italy, Hungary and Romania. The event was successfully organized by the Swiss Geomorphological Society together with the Institute of Geography from the University of Lausanne and supported by the Geosciences/Platform of the Swiss Academy of Sciences but also by cantonal institutions from Ticino (The "Societa Ticinese di Scienze Naturale" and "Museo cantonale di storia naturale"). The hospitable local community of Blenio (commune di Blenio) had also an important role.

A number of 27 presentations, including posters were selected by the scientific committee for the programme. Their thematic was centered on the alpine and mountain regions morphodynamics' under the conditions of the environmental changes. Some contributions must be noticed: the morphodynamics of rock glaciers in the Swiss Alps (R. Delaoye, C. Lambiel and C. Scapozza, E. Cossart and C. Perrier), the morphodynamics of rock slopes after the End of the Little Ice Age in the Northern French Alps (L. Ravanel and P. Deline), the geomorphosites inventory, analysis and touristic integration in the Swiss Alps and Jura Mountains (B.Maillard, A. Perret, S. Morard and A.Staub), the high magnitude torrential processes in the Swiss Alps (C.Graf). An interesting presentation focused on the implementation of the international mountain route of "Via Geo Alpina" which will cross some of the most important areas with geomorphosites in the Alpine chain (M.Schlup et al.).

Other contributions focused on new research techniques applied in mountain landscape and morphodynamics investigation like the Synthetic Aperture Radar interferometry (T. Strozzi and R. Delaoye, S. Mari et al.), the dendrogeomorphological analysis (C. Scapozza et al., V. Garvaglia et al., L. Astrade et al.), the GIS and remote sensing mapping and modeling of the morphodynamic features (G. Toth, Mihai et. al etc.).

A poster section was focused on the young geomorphologist research output but not only. There were included original contributions like the Swiss geomorphosite inventory (S. Martin), the debris slopes features in the Swiss Alps (D.Abbett et al., C. Dvorak et al.) the didactics of geomorphology (S. Morard et al.), the morphodynamics of torrential systems (D. Theler et al.), the mountain landscape changes (L.Laigre et al.) or some features regarding the local features of rock glaciers (G. Ramelli et al.). A really interesting presentation was focused on the water supply systems of "bisses" from the Swiss canton of Valais, in the Bernese Alps, made by prof. dr. E. Reynard from University of Lausanne, a specialist of these issues.

The meeting was a complete one with a new volume presentation (C. Scapozza and G. Fontana), focused on the local geomorphic features (Le Alpi Bleniesi: storia glaciale e patrimonio geomorfologico) and an introduction in the study of geomorphology, a conference of a local researcher (M. Antognini), from the Cantonal Museum of Natural History in Lugano.

The last day a field trip was organized around the area of Lucomagno/Lukmanier Pass (1916 m) and along the Valley of Santa Maria. The main organizer, Cristian Scapozza and some collaborators choosed an interesting walking route in an area with a complex geology, a region rich in periglacial morphodynamics and karstic morphology. There were discussed issues of applied geology and geomorphology (the area is crossed by an old transalpine road and will be crossed in the near

future, in the underground, by the Gotthard Basis Tunnel), glacial and periglacial geomorphology, karstic evolution and environmental protection of alpine river floodplain areas.

The conference was a real opportunity to be in contact with some of the latest achievements of the Swiss geomorphologic community. The selection of a wonderful mountain valley and town in Ticino and the integration of science with the local hospitality were perfectly merged by Cristian Scapozza, an enthusiastic young geomorphologist

born in this area, and by Professor Emmanuel Reynard our host from the Institute of Geography of the University of Lausanne and his team.

The Romanian Association of Geomorphologists was represented by prof.dr. Bogdan Mihai. He received the invitation of the Swiss Geomorphological Society and the Institute of Geography of the University of Lausanne to join the conference with a financial support of the Swiss SCOPES programme.



Prof. dr. Bogdan MIHAI

