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Geomorphic characteristics of dryland distributive systems

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ABSTRACT

Dryland distributive systems include different types of depositional bodies: distributary systems, alluvial fans, fan deltas, lacustrine deltas and sea deltas. Geomorphic data of 81 distributive systems from different arid lands in the world were selected and simple geomorphic parameters such as distributive system area, the average length of the radius, average slope and apex angle were measured. The data analysis indicates that these typologies have different characteristics, though the most distinctive groups include fans, deltas and distributary systems, the geomorphic characteristics of which are constantly intermediate between the two former groups. The results seem to confirm the hypothesis of De Chant et al. (1999) that the shape of distributive systems is mainly controlled by sediment diffusion, channel processes and accommodation space characteristics.

KEYWORDS distributive systems; ephemeral streams; fans; deltas; drylands

1. Introduction

All rivers have an upstream start, in the headwaters, and a downstream end when they reach their base level, be it a water body (sea, lake, or another river) or the lower elevated land they are not able to cross. While erosion is the main process in the headwaters, deposition dominates in the river terminus (Schumm, 1977). Here, commonly in association with a sharp change in bottom gradient and/or some frontal confinement (e.g., the water body or a flat land), stream power substantially decreases and river channels become transport limited as sediment transport capacity is exceeded by sediment supply,

thus resulting in sediment accumulation. High depositional rates result in bank crevasses, channel bifurcation and avulsions (Bristow et al., 1999; Jones and Schumm, 1999; Charlton, 2008; Donselaar et al., 2013). The occurrence of all these fluvial processes give way to alluvial fans and inland distributary systems and, when combined with wave action and longshore drift, lacustrine deltas and fan deltas result.

Several papers have been published on the geomorphology of alluvial fans (e.g., North and Warwick, 2007; Harvey, 2011 and Ventra and Clarke, 2018, for a comprehensive review) and sea deltas (e.g., Galloway, 1975; Bhattacharya, 2006; Olariu and

Bhattacharya, 2006; Syvitski and Saito, 2007; Woodroffe and Saito, 2011; Mikhailova, 2015). Depositional features, stratigraphy and facies of distributive systems have also attracted the attention of scientists also for their potential as oil reservoirs (Kelly and Olsen, 1993; Fisher et al., 2007; Nichols and Fisher, 2007).

Most of the studies on alluvial fans include examples from different climates, whereas those on deltas commonly deal with perennial rivers. Inland distributive systems are typically found in arid regions and make up the terminus of ephemeral streams (Mukerji, 1976; Billi, 2007). In arid regions, with the exception of very steep fans, distributive systems are important because they are among the very few land portions on which agriculture is possible due to the flat morphology, the occurrence of fertile soils and occasional surface and subsurface water supply from the rivers distributary channels that can guarantee some soil humidity. The understanding of dryland distributive systems geometry is also important in oil exploration since no, even conceptual, models have been proposed yet. Therefore, this paper is on distributive systems formed by ephemeral streams in arid lands. Ephemeral streams are dry most of the time and have water flowing only occasionally, following sporadic intense rainstorms. The intermittence of water flow, often characterized by transport capacity limited conditions, makes these depositional systems unique and though they are an important geomorphological aspect of arid environments, they are still very poorly known.

North and Warwick (2007) dispute terms like "terminal-fan" or "fluvial distributary system" because they find a discrepancy between the depositional models adopted by sedimentologists and the modern analogs. Yet, these authors consider distributary systems as relict lacustrine deltas, influenced by non-fluvial processes, though they provide only sedimentological evidence in support of their hypothesis.

The aim of this paper is to contribute to clarifying such uncertainties by means of geomorphic characterization of arid region ephemeral stream distributive systems, which are very poorly known as distinct climatic endmember, notwithstanding the

comprehensive and inspiring work by Hartley et al. (2010), which, however, is mainly aimed at investigating geotectonic control on modern fluvial distributive systems and at providing a descriptive database for comparison with the rock record, though no distinction is made between the distributive system of different climates.

2. Distributive systems terminology

North and Warwick (2007) proposed an interesting and stimulating, though not exhaustive, discussion about distributive systems terminology and Hartley et al. (2010) identified six broad plan form types, based mainly on channel number, sinuosity and bifurcation. For consistency with past and future studies, a simplified classification of ephemeral stream distributive systems is adopted in this study on the basis of previous work (e.g., Stainstreet and McCarthy, 1993; Billi, 2007). In this paper, the term "distributive" is used as a higher hierarchical definition including all the more or less flat cone-shaped depositional bodies formed by a single channel of any sinuosity lateral shifting, though keeping hinged on the distributive system apex, or multiple channels of any sinuosity splitting from the main river stem at the apex or within the main body of the distributive systems.

In arid regions, five main types of distributive systems were identified: distributary systems *sensu stricto*, alluvial fans, fan deltas, lacustrine deltas and sea deltas (Fig. 1). The distributary systems typically originate from flow splitting and the formation of few distributary channels of smaller size than the main river stem, which in their turn bifurcate as they proceed downstream (Fig. 1a). Bifurcation ends where flow reduces substantially for infiltration, channel width decreases from hundreds of meters to centimeters and water vanishes into shallow puddles (Billi, 2007). Distributary systems are not confined and are commonly found on the bottom of closed basins. Their sediment is predominantly sand with small percentages of gravel and scattered boulders, even of very large size.

Alluvial fans form at the exit of a narrow valley at the downstream end of the mountainous part of the catchment. Due to the vicinity to the sediment supply area and a sharp change in gradient, sedimentation

rates are very high and the depositional body may be very thick. Typically, several channels originate from the main river stem in the fan apex, but also one individual channel may shift across the fan, thus contributing to its construction. Alluvial fans may not be the river terminus as the stream channel may proceed beyond it also for many kilometers (Fig. 1b) to reach its main base level. Alluvial fans may be laterally confined, especially for the growth of adjacent fans, whereas they are seldom frontally confined, but when the fan is formed at the confluence with a larger river in a narrow valley. Fan deltas have the same characteristics as alluvial fans,

but the former are always frontally, partially confined by a water body (sea or lake) (Fig. 1c).

Lacustrine and sea deltas are very similar (Fig. 1d and e), but for a few differences: wave action on the former is almost negligible, whereas, in the latter, waves may play a crucial role in controlling the delta size. In arid regions, sea deltas are very rare because most ephemeral streams very seldom reach the coast during exceptionally large floods. The infrequent floods of ephemeral streams result in a low sediment supply to the coast and the small amount of sediment accumulated can be easily dispersed by the waves and the longshore drift.



Figure 1 The main typology of distributive systems considered in this study: a) distributary system, *sensu stricto*; b) alluvial fan; c) fan delta; d) lacustrine delta; e) sea delta; f) the main geomorphic parameters measured: area is the area of the distributive system body; R_d is the radius; α is the apex angle.

3. Data and methods

Through inspection of the planet's arid lands by means of Google Earth Pro®, 81 ephemeral streams distributive systems were selected (Fig. 2). Distributive systems of allogenic, perennial, or seasonal rivers crossing arid regions were not considered for the high annual precipitation in the far headwaters that distinguish these rivers from ephemeral streams, which are dry for most of the year and, in a few cases, also for several years in a row, and have water flowing in the channel only in response to sporadic heavy rainfalls. Also, distributive systems naturally or artificially laterally

confined or affected by human installations or infrastructures were not considered.

For each distributive system, the following parameters were measured on Google Earth Pro® images: distributive system area (A), average length of radius (R_d), average slope (S) and apex angle (α) (Fig. 1f). A minimum number of three radii were traced on the distributive body (Fig. 1f) and their length averaged to obtain R_d . The distributive system slope (S) is the average of the slopes of the individual radii traced to calculate R_d and the apex angle was measured after tracing the distributary system external borders (the dotted line in Figure 1f). A summary of the data measured is reported in Table 1.

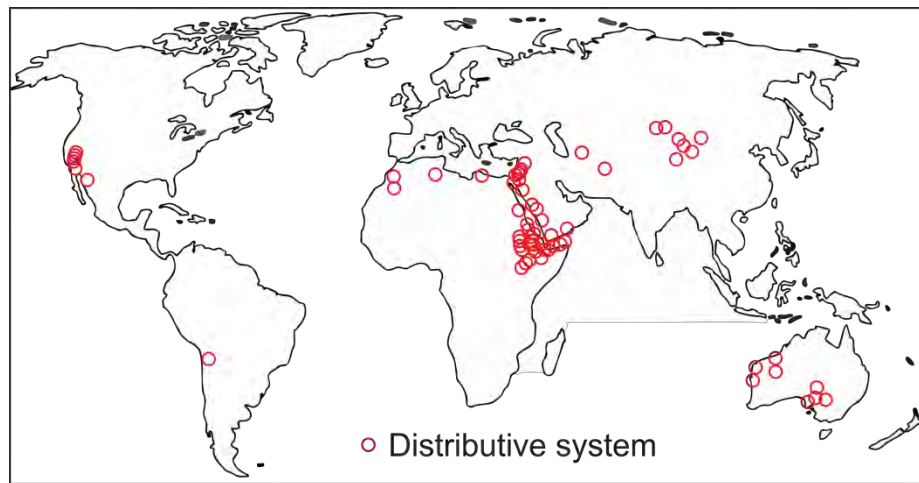


Figure 2 Location map of the distributive systems selected

In this study, the upstream catchment area was not considered for the following reasons:

i) Several papers dealt with the control of catchment area on fan size and slope (e.g., Blair and McPherson, 1994; Viseras et al., 2003; Harvey, 2011; Mokarram et al., 2014) and several power functions with different coefficients and exponents were presented. Their general forms are:

$$A = cA_c^b, \quad (1)$$

$$S = qA_c^k, \quad (2)$$

in which A is fan area, A_c is catchment area, S is fan slope, c and q constants and b and k exponents. According to the extensive work of Harvey (2011), who considered several published data of dryland alluvial fans, c varies between 0.1 and 2.1, b varies between 0.7 and 1.1, q between 0.03 and 0.17 and k

between -0.15 and -0.35. However, if we consider other studies (e.g., Marchi and Tecca, 1995; Viseras et al., 2003; Mokarram et al., 2014; Zhang and Lu, 2014), the ranges are even wider: $c = 0.08 - 3.83$; $b = 0.11 - 3.33$; $q = 0.03 - 1.90$; $k = -0.10 - -0.53$. Harvey (2011) attributes such a large variability to regional differences in fan age, catchment geology and sedimentary processes. For Nevada and California fans, in fact, Lecce (1991) found an inverse relationship between basin erodibility and fan size.

ii) In their study on distributive fluvial systems in the drylands of northern Patagonia, Bilmes and Veiga (2016) found that the catchment area is not the only factor controlling the size of the distributive systems as the accommodation space may also play an important role. The role of accommodation space is also emphasized by Hartley et al. (2010). Viseras et

al. (2003) found that in southern Spain there is no relation between fan slope and catchment area, under conditions of base-level rise.

iii) De Chant et al. (1999) concluded that the fan morphology is predominantly controlled by channel processes rather than environmental variables (climate, lithology). Their model deliberately does not include the sediment influx (which is very difficult to quantify) and it is synthesized by the following equation:

$$V = 0.28[(\alpha/360)\pi R_d^2 h] , \quad (3)$$

in which V = fan volume; α = fan apex angle (Fig. 1e); R_d = the fan radius; h = the total height at the fan head with respect to the fan fringe.

iv) With the exception of estuarine deltas, the size and shape of sea deltas are largely controlled by the action of the waves, the distance of the river basin mountain portion from the coast, the depth of the nearshore water and the tide range (Woodroffe and Saito, 2011). Though for lacustrine deltas, the relationship between catchment size and delta area seems obvious, for sea deltas, it is not so straightforward.

Given the limited control that catchment area may or may not have on the area of distributive systems, it seems preferable not to consider this parameter in the attempt to discriminate the different types of distributive systems mainly on the basis of geomorphic parameters. Moreover, the role of the catchment area has already been widely investigated for alluvial fans, though with large uncertainties and often with unsatisfactory correlation coefficients, the investigation of which is beyond the scope of this study. Following the results of De Chant et al. (1999) and the work of Viseras et al. (2003), it is evident that the shape of the distributive system is largely influenced by the depositional processes and the accommodation space characteristics (Hartley et al., 2010), rather than the catchment characteristics. For these reasons, the characterization and discrimination among the different types of distributive systems were carried out through the analysis of the four parameters, defined above and outlined in figure 1f, that can be used to describe the geometry of any type of distributive system.

4. Results

Mean and range values of the distributive system geometry parameters are reported in Table 1. From these data, some main preliminary differences among the types of distributive systems are evident. For instance, lacustrine and sea deltas have the smallest slopes, whereas the distributary systems have the smallest apex angle and the lacustrine deltas the widest.

One of the main aims of this study was to verify if distributary systems are simply relict lacustrine deltas as proposed by a few authors (e.g., North and Warwick, 2007), rather than a depositional body with distinctive geomorphic characteristics. The plot diagram of distributive area versus slope (Fig. 3a) clearly shows a distinction between inland distributary systems and lacustrine deltas, with the latter and the former data plotting above and below, respectively of the discriminating function, expressed by the following power equation:

$$S = 0.0109 \times A^{0.508} , \quad (4)$$

Additional evidence of the geomorphic differences between distributary systems and lacustrine deltas was obtained by bivariate diagrams considering the combination of the distributary radius, slope and area vs. the apex angle (Fig. 3b, c and d). In Figure 3b to d, however, the discriminating function is not derived by correlation analysis, as in figure 3a, but it is traced by eye and its aim is simply to emphasize the distinction between the two data sets that plot in different areas of the diagrams.

The data of Table 1 indicate that both alluvial fans and fan deltas have very similar characteristics in terms of slope and area, but they are well distinguished from lacustrine and sea deltas. This observation is also confirmed by the diagram of Figure 4 in which the distributive system area is plotted against the slope. In Figure 4, in fact, alluvial fan and fan deltas are very close and take up the upper portion of the diagram, whereas lacustrine and sea deltas are mixed up and stand in the lower portion with distributary systems in an intermediate position. The constant, the exponent and the correlation coefficient of the equation,

$$S = qA^k , \quad (5)$$

are reported in Table 2.

The k exponent of alluvial fans, fan deltas and lacustrine deltas is very similar, whereas that of

distributary systems and sea deltas are the largest and lowest, respectively.

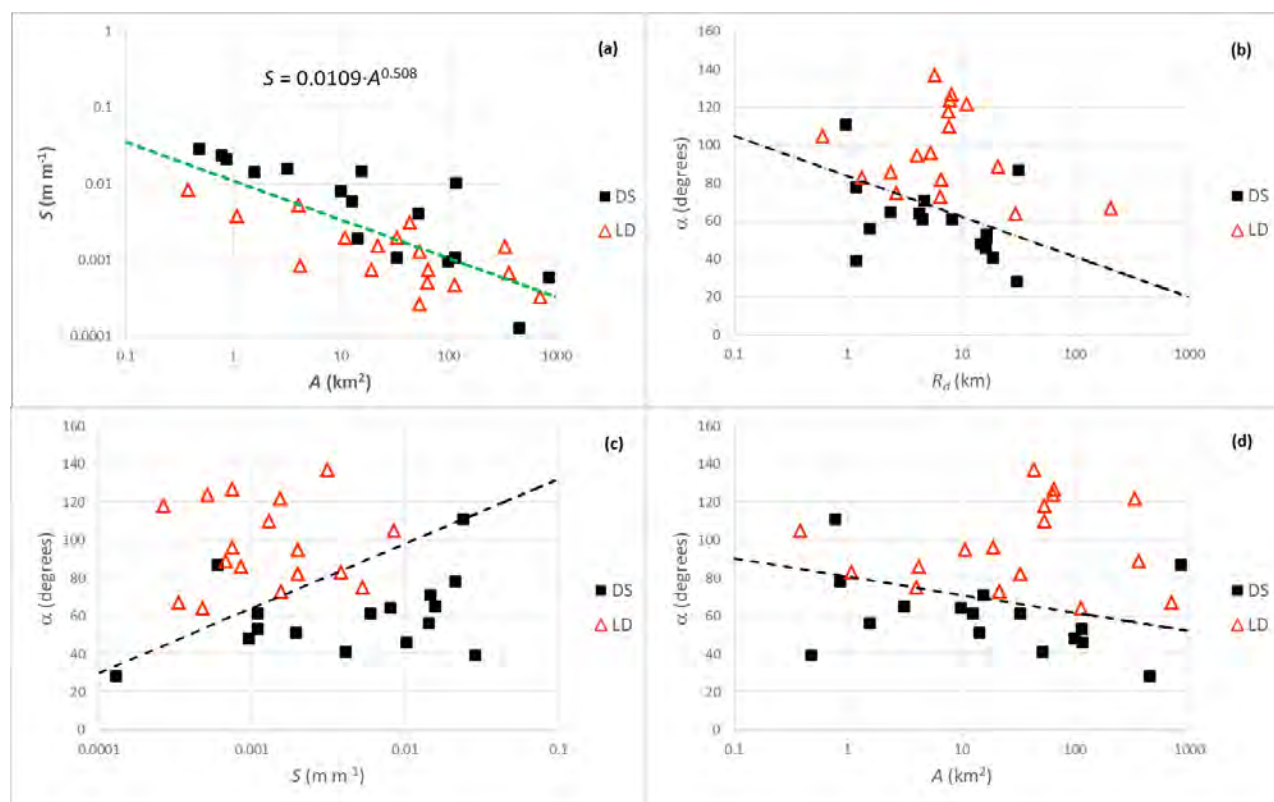


Figure 3 Bivariate diagrams to discriminate between distributary systems (DS) and lacustrine deltas (LD): a) distributive system area (A) vs. slope (S); b) radius (R_d) vs. apex angle (α); c) slope (S) vs. apex angle (α); d) area (A) vs. apex angle (α)

Table 1 Summary of the geomorphic parameters measured: A = area; R_d = radius; S = slope; α = apex angle (see Fig. 1f); cv = coefficient of variation

		Distributary systems	Alluvial fans	Fan deltas	Lacustrine deltas	Sea deltas
A (km^2)	mean	110.90	374.79	100.00	110.14	279.46
	max	852.34	2176.29	1273.37	700.47	2157.24
	min	0.48	1.58	0.02	0.38	0.26
	cv	0.49	0.54	0.31	0.58	0.47
R_d (km)	mean	10.92	14.87	6.21	19.46	12.11
	max	31.88	50.42	45.00	203.00	56.38
	min	0.96	1.51	0.18	0.60	0.58
	cv	1.06	0.90	0.50	0.39	0.72
S (m m^{-1})	mean	0.0095	0.0350	0.0687	0.0020	0.0042
	max	0.0288	0.1090	0.1833	0.00845	0.0119
	min	0.0001	0.0051	0.0024	0.0003	0.0005
	cv	1.03	1.08	1.10	0.89	1.20
α (degrees)	mean	60	86	79	97	81
	max	111	121	108	137	126
	min	28	32	36	64	38
	cv	2.97	3.56	4.10	4.23	3.53

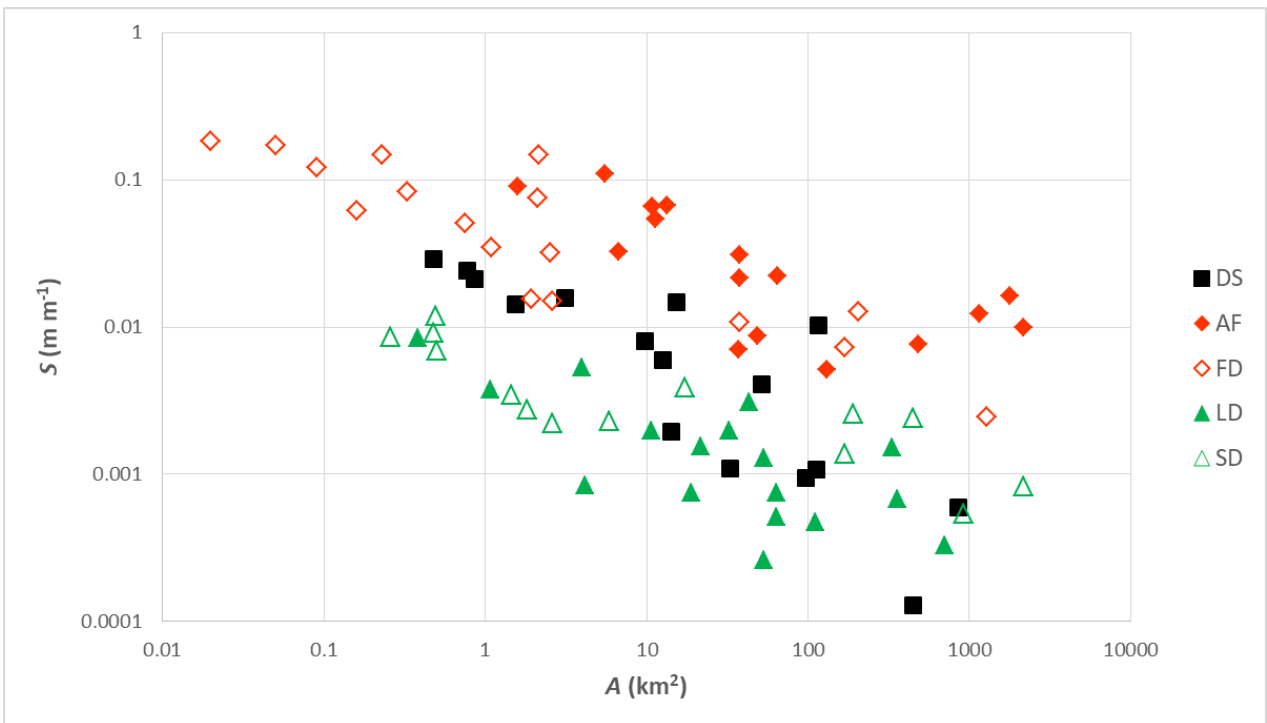


Figure 4 Plot diagram of distributive area (A) vs slope (S). The five types of distributive systems are clearly distinguished; DS = distributary systems; AF = alluvial fans; FD = fan deltas; LD = lacustrine deltas; SD = sea deltas

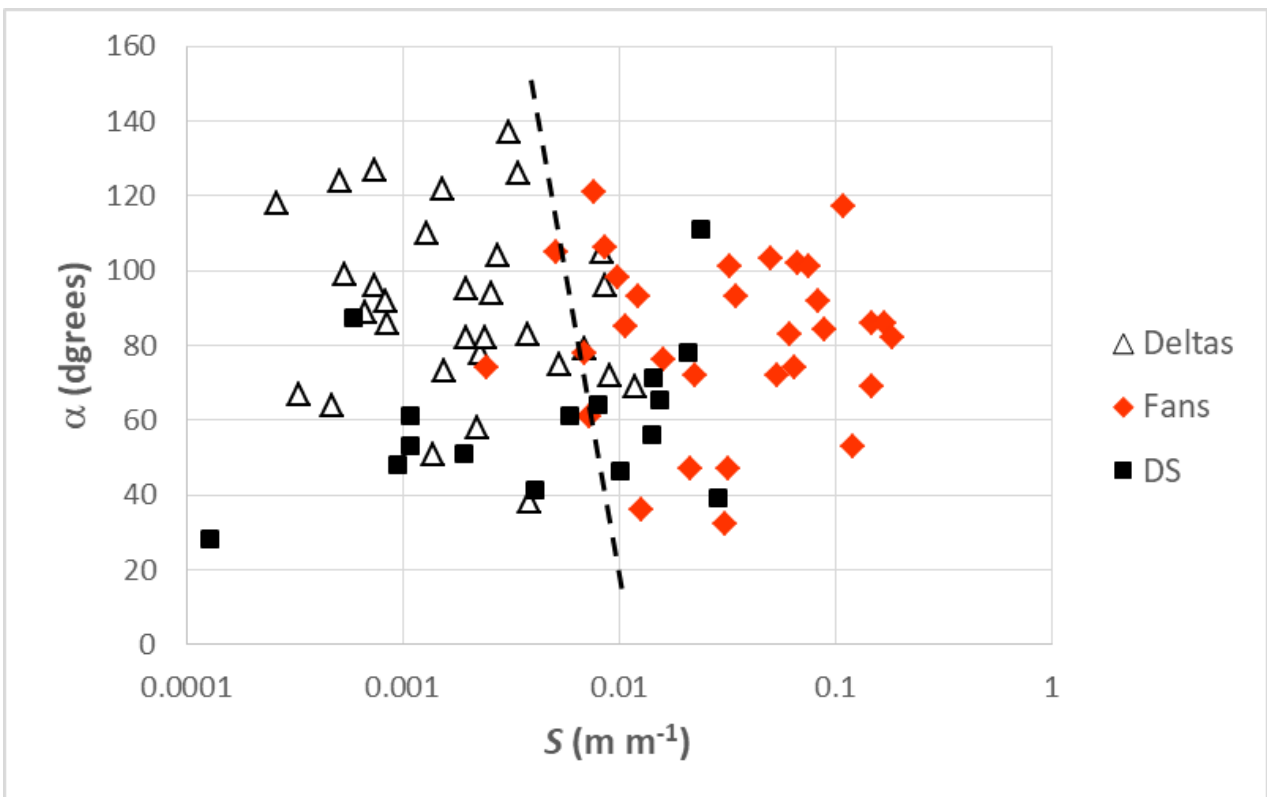


Figure 5 Plot diagram of distributive system slope (S) and apex angle (α). Deltas and fans take up different parts of the diagram. The discriminatory line (dashed line) is not an interpolation line and it is traced by eye

Other non-spurious correlations, such as that between slope and apex angle, are not significant, but if we group the data into the three main types of distributive systems, i.e., deltas, distributary systems and fans, a clear distinction is evident between deltas and fans, whereas distributary systems, yet occupy and intermediate position (Fig. 5).

Table 2 Constant, exponent and correlation coefficient of equation (5) (DS = distributary systems; AF = alluvial fans; FD = fan deltas; LD lacustrine deltas; SD = sea deltas)

Dist. Syst.	q	k	R^2
DS	0.0220	-0.588	0.72
AF	0.0845	-0.329	0.52
FD	0.0490	-0.370	0.80
LD	0.0040	-0.347	0.51
SD	0.0054	-0.241	0.72

The shape of a distributive system depends on many factors, but sediment diffusion and channel processes are probably among the most important, as suggested by De Chant et al. (1999). In fact, if no later constraint is present, the rate of channel shifting, avulsions and bifurcation are expected to result in wider and gentler distributive systems.

To explore this issue, a cone shape parameter $Y = \alpha \times S$ is introduced. The plot diagram of area and Y (Fig. 6) clearly shows that distributary systems, fans and deltas have different shapes. The respective interpolating functions are:

- Deltas: $Y = 112.2 \times A^{1.069}$, ($R^2 = 0.94$) (6)

- Fans: $Y = 1391 \times A^{0.2868}$, ($R^2 = 0.59$) (7)

- Distributary systems: $Y = 2869 \times A^{0.5404}$, ($R^2 = 0.69$) (8)

Small deltas have the lower values of the $\alpha \times S$ product, i.e., narrow and flat bodies, but large deltas have the largest $\alpha \times S$ product and, since an increase in delta gradient is not realistic and increase in apex angle has to be expected. Small distributary systems

have the largest values of the $\alpha \times S$ product implying comparatively steeper body slope, but for large distributive areas are similar to the deltas. Fans have an intermediate behavior with a lower rate of change with increasing the fan area.

5. Discussion

Hartley et al. (2010) have shown that the geometry of inland distributive systems is controlled by a number of parameters, the most important of which are climate, discharge, sediment supply and accommodation space. This study focused only on dryland rivers, therefore, the climate variable can be neglected. In this study, however, also lacustrine and sea deltas were considered as part of the family of distributive depositional bodies. Deltas are also controlled by discharge, sediment and accommodation space, but especially the geometry of sea deltas and fan deltas is also influenced by the wave action. Furthermore, the fluvial and depositional processes are different for distributary systems, alluvial fans, fan deltas and deltas and the confined size characteristics of the deposited material may also play an important role. Notwithstanding such a complex interaction of factors, acting at different rates and with different intensities, yet it is possible to discriminate the geometry of the main types of distributive systems on the base of simple geomorphic parameters (Figs. 4, 5 and 6). Hartley et al. (2010) have produced a huge, planet-wide data set on large distributive fluvial systems, but they recognized six main planforms types based on the number, morphology and downstream changes of the river channels, thus stressing the role of fluvial processes in determining the distributive system characteristics. However, Hartley et al. (2010) included in their analysis any kind of distributary system that corresponded to their definition of a "deposit of a fluvial system which in planform displays a radial, distributive channel pattern". In this regard, this study, though restricted only to arid land distributary systems and based on a smaller data set, may be seen as a partial development of Hartley et al. (2010) work.

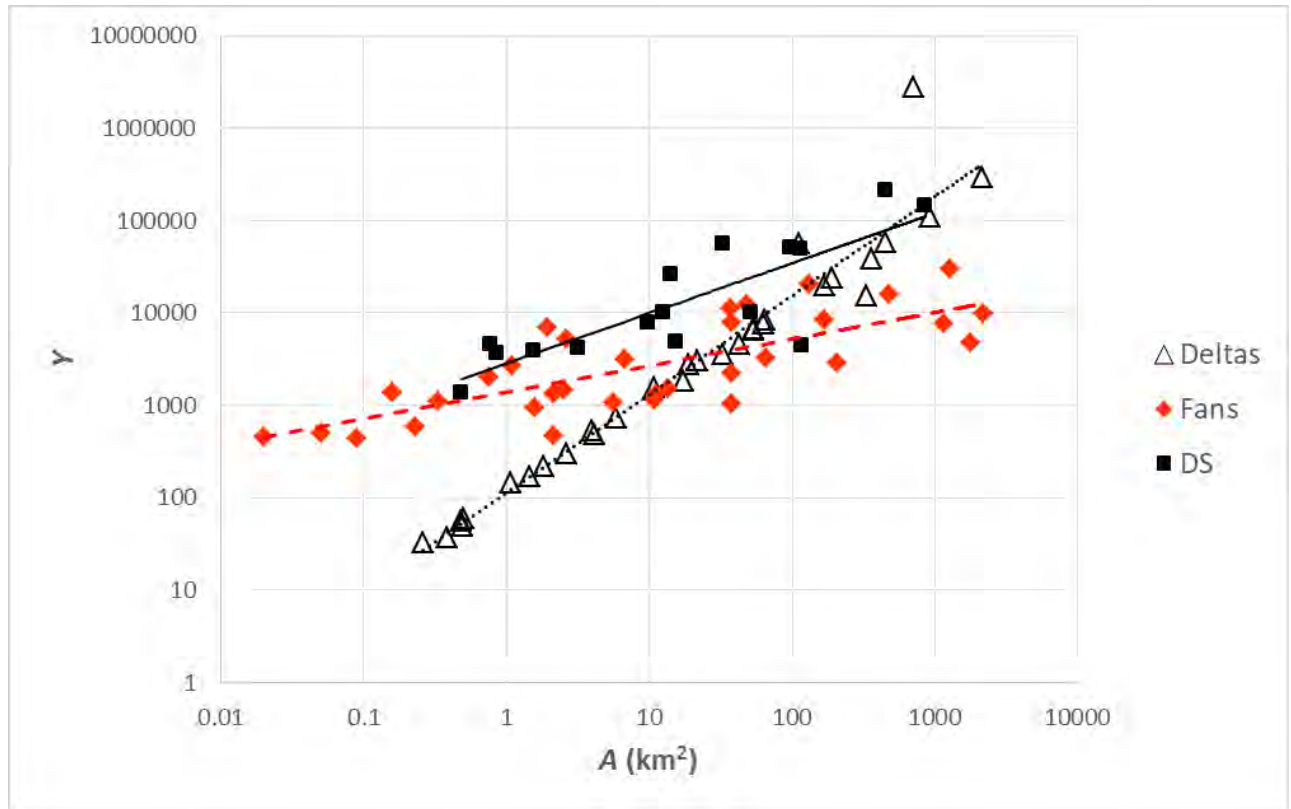


Figure 6 Plot diagram of distributive system area (A) vs. a distributive geometry parameter $Y = \alpha \times S$, where α is the apex angle and S is the slope

A few authors (e.g., North and Warwick, 2007) are convinced that distributary systems are lacustrine deltas inherited from wetter times. According to these authors, modern distributary systems were originally lacustrine deltas and, as the lake dried up, they became entirely subaerial and maintained the distributive geomorphology. This interpretation is based on the occurrence of lacustrine deposits interbedded with ephemeral stream deposits. Though the lacustrine distributary system model evolution may find some correspondence in the world, in many arid and semiarid endorheic, structural basins distributary systems, they have no relation with a former lacustrine delta simply because no lacustrine deposit is present and no lake or swamp ever formed in the past. This setting is quite evident in the Raya graben in northern Ethiopia, where almost all rivers form distributary systems, but no lacustrine deposit was identified by both surface and subsurface drilling investigations (Ethiopian Government, 1977; Billi, 2007).

In the present study, in the lacustrine deltas category, deltas of former lakes, now desiccated, were included as well, but the diagrams of Figure 3 clearly show that distributary systems and lacustrine deltas occupy separate parts of the plot area, with very limited overlapping. In fact, for a comparatively similar area, distributary systems have a steeper slope than lacustrine deltas and for comparatively similar radius length, lacustrine deltas have a much wider apex angle. In other words, distributary systems are steeper and narrower than lacustrine deltas. Such a difference can be accounted for by the different depositional processes that, in the case of the delta, are influenced by the frontal constraint of the water body, whereas in the distributary systems, channels are free to prograde, provided their water and sediment discharge is properly sustained from upstream. Moreover, seasonal or inter annual lake level fluctuations may favour the lateral, longshore redistribution of sediment, resulting in a wider apex angle.

The wave action seems to have less influence on fan deltas, probably because they are made up of coarser material. Alluvial fans and fan deltas have very similar geomorphic characteristics as do lacustrine and sea deltas, though these two groups of distributive systems are clearly distinct. Deltas have comparatively lower slopes than fans (alluvial and fan deltas) but comparable apex angles. This likely implies higher depositional rates at the fan apex, but also some influence of apex bedrock topography and tectonic displacement effect cannot be excluded.

Distributary systems appear to have intermediate characteristics between fans and deltas, but they maintain their unicity in terms of fluvial processes and geomorphic characteristics. Bifurcation and avulsion are common channel processes in distributary systems, with main flood flows often shifting from one distributary channel to the other, but multiple active channels, like those observed on fans, are absent and in deltas, the rate of bifurcation is higher. Deltas, in fact, have the lowest gradient and, though the bedload flux may be less than in distributary systems, the frontal confinement imposed by the water body and the smaller stream power may result in high sedimentation rates and, consequently, in high bifurcation rate.

The prevalence of factors such as accommodation space characteristics, fluvial processes and sediment supply in shaping distributive systems was pointed out by Hartley et al. (2010), who also stressed the importance of local geological and structural settings. The clear distinction of the different kinds of distributive systems highlighted in this study by means of bivariate diagrams including only geomorphic parameters seems to support the conclusions of Hartley et al. (2010) and the approach presented by De Chant et al. (1999). These authors derived a simple equation to calculate the volume of alluvial fans as a function of the apex angle (α), the radius (R_d) and the elevation of the apex with respect to the fan fringe (h) (eq. 3 of this paper). In the present study, it was not possible to measure h , which would require extensive field inspection to determine the elevation of the bedrock under the fan apex and the main focus was instead on distributive system area. Moreover,

distributary systems and deltas have lower gradients than fans and allow to remove h from eq. 3 and to define a geometry parameter Y as the product of apex angle and slope. The plot of area vs the shape parameter Y , confirms that distributary systems, deltas and fans have distinctive shapes (Fig. 6). Deltas and fans have the highest and the lowest correlation coefficient ($R^2 = 0.94$ and 0.59 , respectively), probably because they are the flattest and have the thickest bodies, respectively. Unlike deltas and fans, the distributary systems of this study are all unconfined and this may be an important factor of distinction. Nevertheless, these results indicate that for fans h is an important parameter and indirectly confirms the approach of De Chant et al. (1999).

6. Conclusions

A total of 81 ephemeral streams distributive systems were selected from different arid lands in the world. They were included into five categories: distributary systems *sensu stricto*, alluvial fans, fan deltas, lacustrine deltas and sea deltas. The data show that distributary systems have geomorphic characteristics different from those of lacustrine deltas and no relations seem to exist with former lacustrine deltas of wetter climate as instead advocated by North and Warwick (2007).

Alluvial fans and fan deltas have similar fan geometry and very similar are also the geomorphic characteristics of lacustrine and sea deltas. Distributary systems have intermediate geomorphic characteristics between fans and deltas. Fan deltas seem to be less influenced by wave action than deltas, likely because the former are composed of coarser sediment. Distributary systems have a smaller apex angle and are steeper than deltas. The lower gradient of the latter favours the formation of multiple channels and hence a wider apex angle. All these results confirm the convincement of De Chant et al. (1999) that the shape of distributive systems is mainly controlled by sediment diffusion, channel processes and accommodation space characteristics rather than catchment area.

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Geomorphological restitutions for the geomorphological regionalization of Romania: the Moldavian Plain case

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ABSTRACT

In Romanian geomorphology, there are a series of inadvertencies induced socio-politically by the oppressive regime before 1990. Among these, we exemplify in this paper the use of the notion of plain for an area in Romania, which although exhibits characteristics of a hilly area, both geomorphologically and geomorphometrically: the Moldavian Plain. To support this position, we applied the Hammond and Iwahashi and Pike landform types classifications on the global GMTED DEM for Romania. We added an argumentation of the geology and geomorphological evolution of the study areas together with a history of its geomorphological regionalization. I hypothesize that the communist regime that implemented the collectivization of the arable lands influenced the geographers to introduce some areas in Romania to the plain category to argue for their collectivization. Another hypothesis might be that the inclusion of the plain landform typology increased the allocation of mechanization for agriculture. While the further study might reveal which of the theories is correct, the critical part is the restitution of the valid typology. In this sense, I support the elimination of the plain term to designate the landforms of this region and the reintroduction of the terms used before 1945, like Jijia Depression or of the term Jijia Hills, since the global geomorphometric classifications of landforms support this category.

KEYWORDS geomorphological regionalization; plain; hill; Jijia Hills; Jijia Depression

1. Introduction

In Romanian geomorphology, there are a series of inaccuracies in the regionalization, which we consider that appeared due either to a limited knowledge from the moment of their study or a socio-political context that imposed a particular vision. We refer here to the oppressive regime before 1990, which probably argued the need for

collectivization or for increasing the mechanization of agriculture, thus establishing the criteria for areas that, although they did not fit typologically and geomorphometrically to the "plain" category, had to be declared this way.

The most typical situation of this kind is the plains in the hilly regions (Fig. 1): the Moldavian Plain (Fig. 2) and the Transylvanian Plain. These are neither geomorphological nor geomorphometric plains, but

they have been imposed as a name for other reasons. I will not treat the Transylvanian plain problem, but I will stop at the Moldavian Plain. Initially, I will show the evolution of the knowledge about it and then show the geomorphometrical reality with the help of several classifications of the landforms on geomorphometric bases. An initial study on this problem was made by Niculiță (2020).

I also emit several hypotheses for such evident wrong labeling by showing the historical evolution of the terms used to label this region.

Someone might ask why we need to resolve this issue now since, in the geomorphological literature, regionalization is no longer a key research direction. My answer is twofold. First of all, we need a "restitution" of the scientific reality, which was distorted for a political reason. And second, actually, there are real issues of geomorphological research that pose problems if we keep the distorted reality:

- while an outsider of Romanian geomorphology performs peer-review, it might look unrealistic to study landforms (cuestas) or processes (soil erosion, gully erosion, and landslides) that are not normally associated with a plain area with all the consequences that appear from this;

- the regional extension of some geomorphological units is used, for example, in landslides hazard assessment for susceptibility and risk (Günther et al. 2013, 2014, Wilde et al., 2018; Bălțeanu et al., 2020; Grozavu and Patriche, 2020) so from the geological (lithological and structural) and geomorphological (genesis, evolution, and geomorphometry) point of view, geomorphological regions (sub-divisions) have distinct characteristics from each other that are reflected in the typology, density, age and recent dynamics of landslides and this can have an impact when especially European or global approaches take these regions from the literature, without knowing the reality from the field.

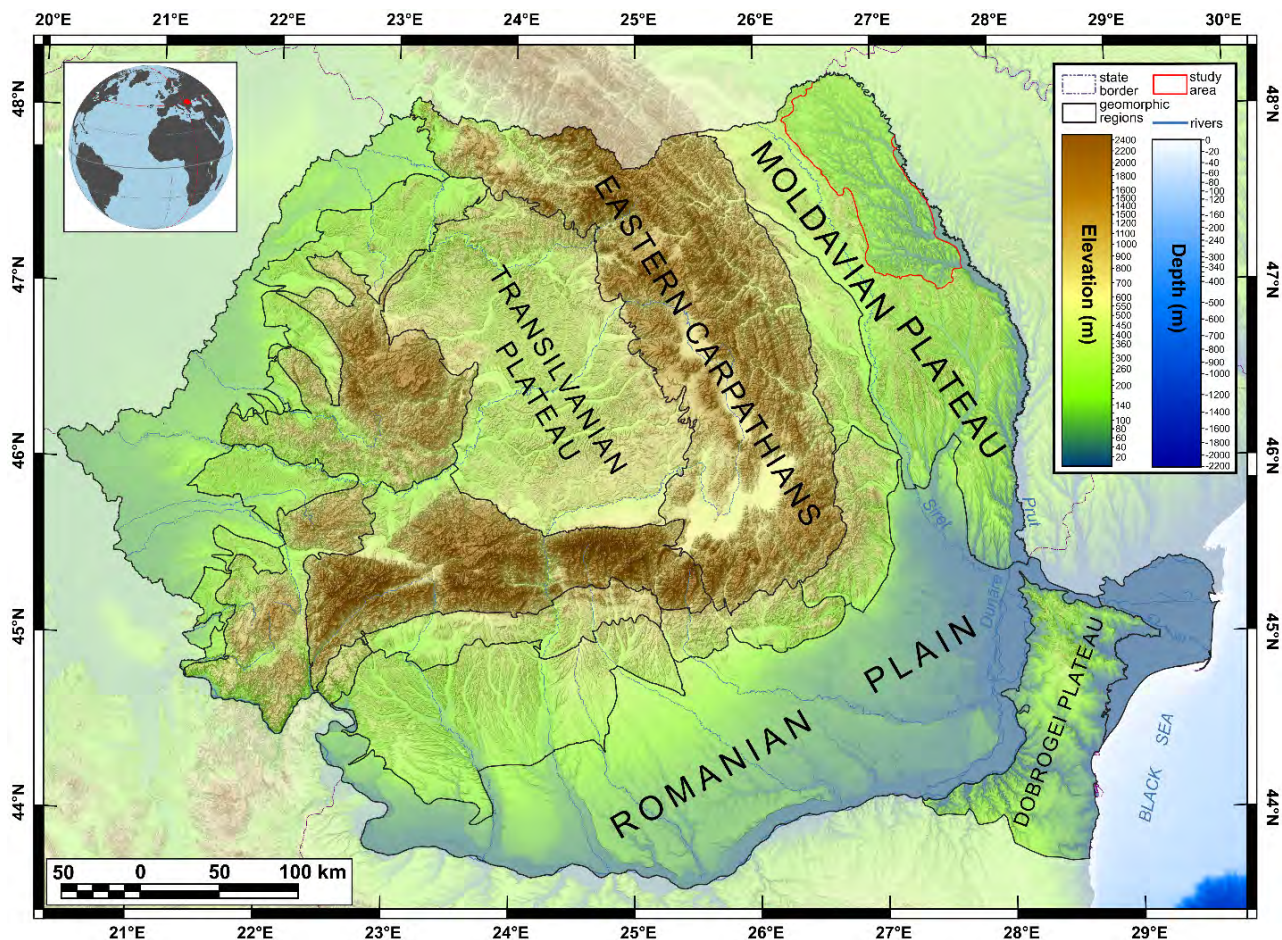


Figure 1 The position of the study area in Romania

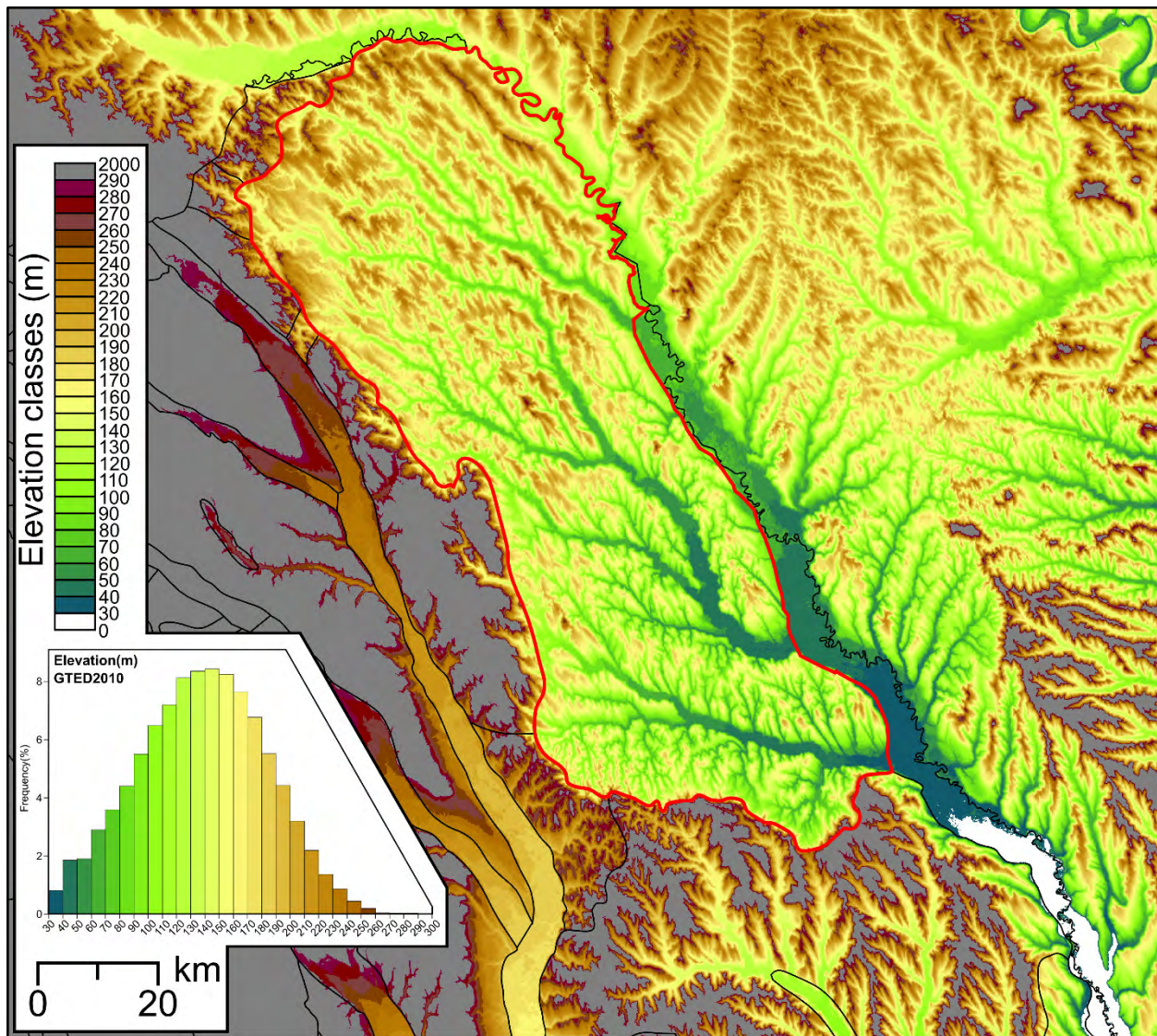


Figure 2 The hypsometry of the study area (GMTED 2010 DEM at 250 m)

2. Methods

2.1 Definitions and differentiations

First, I review the literature regarding the landforms in general and the plains in particular. In the incipient modern geographical literature even (Von Humboldt, 1849, 1856), there is talk of high plains, from the Asian and South American mountain ranges, compared to the low plains, close to sea level, so there is no threshold defined for the absolute altitude of a plain. Penck (1894) dedicates a lot of space to the discussion about plains and their genesis: he points up to the existence of accumulation plains, some of which are coastal, other tectonic, while others can be aggraded by human intervention

through river regulation (Po, Rhin, Mississippi). He also mentions the valley fan plains at the base of mountains. Also, in the context of the Davisian theory of cycles of erosion, the peneplains are considered to have high-altitude plains (Johnson, 1916; Fenneman, 1936), which are also considered "interior plains" (Fenneman, 1917) and categorized sometimes as basins or plateaus. Plains can be both erosional (plains of denudation – Gregory et al., 1910, "East Gulf Coastal Plain ... by gradation" – Fenneman, 1917), depositional ("Snake River Plain ... lava plain" – Fenneman, 1917), and aggradational (alluvial plain - Gregory et al., 1910, "Mississippi Alluvial Plain", "desert aggraded plains" of Nevada – Fenneman, 1917), even abyssal (Dunkerley, 2016), but in no

geomorphological typology, the land use is a criterion. An extensive literature is also dedicated to the marine planation plains (coastal plains) and subaerial denudation plains (Davis, 1896; Johnson, 1916, 1919; Cotton, 1955; King, 1963). Dissected plains with a moderate relief ("Dissected Till Plains" – Fenneman, 1917) are also recognized.

What is a landform in general and how we can define them is still a problem in progress in geomorphology and geomorphometry (Hammond, 1962; Weaver, 1965; Hammond, 1965; Evans, 2012). This applies even to landform which might be easy to grasp geomorphometrically, like mountains (Mark and Smith, 2004 – ontological analysis; Derruau, 1997 and Fairbridge, 1997a, mention relief of over 700 m and massiveness, and separate isolated mountain from mountain ranges; Kapos et al., 2000 use absolute elevation over 2500 m and slope and elevation range below; Meybeck et al., 2001 used relief roughness in m/km, but computed as the range of altitude; Fisher et al., 2004 used curvatures; Sayre et al., 2018 uses Hammond classification extended by Karagulle et al., 2017) which requires additional criteria to be used, like geology and tectonics (Murphy, 1968; Fairbridge, 1997a), climate and vegetation (Gerrard, 1990, which also points out that dissection is important; Slaymaker, 2003; Körner et al., 2017).

In Romanian on the Latin route, the term plain (lat. campos = field, infield) refers to an area without major level differences. In English, the term is self-descriptive ("a large area of flat land"¹). Still, the allusion to the agricultural field in the Romanian language can confuse the untrained people by using the term plain for agricultural fields.

The geomorphologic and physiographic plain is an area with flat shape and with low variations of altitude (Finch and Trewartha, 1942). Penck (1894) used qualitative criteria: "Plains are extended stretches of the land surface in which the height differences between neighboring parts are so small that they are almost entirely beyond our immediate perception," but point also to 0.4% for flat plains and up to 10-20% for inclined plains (up to 9°), the focus being on the lack of relief, even though there is a

general inclination or there is a concave shape. When the valleys incise the plain, it becomes a plateau, and when the valleys incise more than 200 m in the plateau level, it becomes tableland, although the threshold is fuzzy between plains and plateaus. He extends the discussion about the difference between plain and plateau by mentioning the flatness of the ridges. When an area is dissected, so the flatness is removed, and relief over 100 m appears, he considers it to become a hilly area. His discussion points also toward the difference between lowland and upland by mentioning that lowland should not be associated necessarily with plains, although in the areas under 200 m elevation, high relief is not possible: the north German lowlands and the interior of Russia are not plains (the first is a tableland while the second is hilly area) and the Danube/Upper Germany and Podolia are plateaus. Penck shows that using vegetation criteria for separating plains can be problematic. Finch and Trewartha (1942) define a threshold for a plain relative relief of 150 m. Further, by relative relief, plains can be classified as (a) flat plains, with relief under 15 m, (b) undulating plains with relief between 15 and 46 m, (c) rolling plains with relief between 46 and 91 m, and (d) rough dissected plains with relief between 91 and 150 m. The area of a plain has to have an order of magnitude given by its size bigger than ~6200 km² in the hierarchy of Linton (1948) and represents a section in the landform unit hierarchy. Considering this, the plain is an area wider than a floodplain, which is a tract unit, and from a flat, that is a site unit (Linton, 1948, 1951; Savigear, 1965), but smaller than a province (>39000 km²). Considering the size order of relief units proposed by Ahnert (1988), the plain is a mesorelief to macrolief order type of site with widths of at least 10 000 m and an area of at least 100 km². Mescherikov (1968) defines a plain as "an area of land surface featuring small differences in topographic elevation and uniform from the geomorphological point of view" in a strict sense, and gives as a synonym the term "flatland" or "platform plain" in a broader sense as territories with a "combination of plains of various origins" linked to the geotectonic areas of shields and platforms (the

1

<https://www.oxfordlearnersdictionaries.com/definition/>

[english/plain 2#:~:text=%E2%80%8Ba%20large%20area%20of%20flat%20land](#)

North American Plain and the Eastern European Plain).

Flat surfaces, as I have shown, are mentioned for higher elevations, cases in which sometimes are considered as plateaus (convex features with a flat top – Derruau, 1977; "an elevated tract of the comparatively flat or level ground" – Fairbridge, 1997b; "an elevated area of relatively smooth terrain, frequently separated from adjacent areas by steep slopes" – Shaw, 2016) or tablelands (Fairbridge, 1997b). Sometimes, these landforms are considered high plains.

Second, I present the evolution of the terminology for the study area in the Romanian geomorphological literature. The early geologist that studied the geology of the Moldavian Plateau (Cobâlcescu, 1862; Simionescu, 1902; Sevastos, 1903, 1907, 1908, 1911, 1922; David, 1914) use terms like "small" and "big" hills, colines, plateaus and tablelands. David (1919, 1921), describing the Sarmatic Plateau, mention that its neighbor to the north is the Prut "bigger" Depression, an intracolone depression. Martonne (1924) mentions that in the field, the area between Siret and Nistru does not fit the definition of a plateau but rather of hills and large valleys. Porucic (1928) geomorphologically regionalizes two Plains, the Moldavian (on both sides of the Middle Prut river) and the Pontic Plains (the coastal plain of Black Sea between the Chilia branch of Danube and Odessa), in the area east of the Carpathians. Mihăilescu (1929, 1931a,b) uses the term depression and characterizes the landforms as low hills: "a region of low hills (the Middle Prut Depression)". Rick (1931) gives a detailed description of the landforms in the study area, which he names Jijia Depression. He remarks on the hills, their asymmetry with the presence of cuesta² scarps (called in Romanian "coaste"³ a notion used in the popular language for steep hillslopes) and dip slopes, the plateaus from the northern part, the wide

floodplains, and the widespread presence of landslides and soil erosion.

Tufescu (1935, 1937) mentions both the Moldavian Plain and the Jijia-Bahlui Depression. However, sometimes it seems to refer to Porucic's Moldavian Plain, with the Jijia-Bahlui Depression as a subdivision of the previous one. Tufescu is much more interested in the boundaries between this low region and its higher neighbor to the west, the Siret Plateau, than in the criteria for framing in a certain typology of the region itself. Tufescu (1935) mentions that the area's inhabitants call the area "plain"⁴. Later, he returns with an aspect that might seem valid, but is nothing more than the trap of the panoramic method and on the literary style of the description: "Viewed from a dominant point on edge, the relief of the depression appears aged, not far from the stage of peneplain. The gaze rests on wide peaks, beasted in the form of low platters (150-200 m) that are lost in the washed sky like weak pencils on frosted glass." (Tufescu, 1935) also reveals endless surfaces, platforms, levels of erosion based on data from Mihăilescu (1929). Of course, the lack of remote sensing data made the geographer and geomorphologist of the interwar period dependent on the panoramic vision and the topographic map in coarse interpolated level curves, so that the typological framing from a geomorphological point of view was made on the basis of the visual analysis of morphology. The planation surfaces that were described also point to the Davisian perspective of the landform interpretation, so a peneplain, which is "almost a plain", easily can be classified as a plain, without taking into consideration its geomorphometry⁵. So, if in 1935 he considered the term depression as a scientific one, and the term plain as a popular one, Tufescu (1941, 1945) comes back to the problem and argue for the usage of the term of plain, including in the school textbooks, because he is considering that the people are not able to grasp the scientific technicalities of the

² Although the recognition of cuerdas is only given since Tufescu (1945)

³ A detailed discussion is given by Sîrcu (1956)

⁴ It is worth mentioning that through the latin, în Romanian, "câmp" and its derivation of "câmpie" refers to the arable use and not to the flatness as in English

⁵ It is worth mentioning that in the majority of the Romanian geographical literature Davis was cited through its german translated books, so probably "almost a plain" was interpreted as „surely a plain"

depression notion which becomes "artificial". He also argues that geographically the vegetation and land use can be considered in order to name a region, and since the locals use the term "câmp" (which can be translated in English as field), then "antropogeographically" is natural to name the area as a "câmpie" (which can be translated in English as plain).

Simionescu (1937) uses the term intracolone depression probably after David (1921) but do not cite him. He considers this intracolone depression for both sides of river Prut, separating the Jijia Depression to the west and Răut/Bălți Depression to the east. While in the text, he uses depression for both of them, in the map showing the morphologic regions of Moldavia, he mentions in the legend these areas as plains (for the map he mentions Tufescu and Mihăilescu as sources). There is also a note saying it can also be named the Moldavian Plain, being similar to the Transylvanian Plain.

Martiniuc (1955) uses the term Moldavian Plain and shows how we can compute the mean landform depth of fragmentation (a variant of relief used in Romanian geomorphology) on topographic maps based on the Chentsov (1948) method. He also shows a relief map for the Iași District. The method uses the number of contours, valleys and ridges, and echidistance along with several profiles for a certain region. The value computed with this method is too generalized since the values reported for Jijia and Bahlui Plain are only 50 to 70 m, while for the Repedea-Păun Plateau to the south is only 100 to 120 m. I invite the reader to see Figure 2 from Niculiță et al. (2018), where a swath profile (15 km wide) for the same area shows relief values over 150 m for the Jijia-Bahlui over 300 m for Repedea-Păun.

Sîrcu (1956) discusses the problem of the tectonic and geologic genesis of the depression, stating the lack of tectonic and structural control and proposing the lithology and river network control on

the quaternary lowering of the study area with rates higher than the areas that have a protective caprock of arenitic and biomicritic limestone facies. Sîrcu is the single researcher that opposes the usage of the term plain for the study area: "far from representing a plain ... the use of either the term plain or hilly plain is nonsense, by plain is meant a region with flat and cumulative relief⁶ and not a region of hills, whatever they may be of age".

Coteț and Martiniuc (1960) use the term of coline⁷ plains (hilly plains), mentioning the verbal information given by L.G. Kamanin (a Russian geomorphologist that participated in the Romanian-soviet field trips from 1955 and 1956) for areas like Liteni Depression, the area of the Moldavian Plateau south of Central Moldavian Plateau and Jijia Plain which together with Bălți Plain is included in the Erosional Plain of the Middle Prut. These authors acknowledge the relief values up to 200 m (considered maximum values) in these areas and mention the Jijia Plain mean relief values of 70 m. In the Figure 40 of their paper, the caption is saying: "Landslides in the Hilly <<Plain>>⁸ of Transilvania with the aspect of mud torrents". In the main landform types of Romania section of their paper, the Jijia type of Hills and Plateaus from platforms is considered to contain low hills and hilly plains, with very active landslides. Its subdivision is the Săveni erosional-structural Plain and the Iași sculptural-accumulation⁹ Plain. Further, Demidovici et al. (1960), in the physio-geographic regionalization of Romania, uses this statement: "The forest-steppe land of the Jijia Depression, known as the Plain of Moldova".

I have to mention here that after 1945 in the Romanian geomorphological literature, there was an infusion of soviet literature. The most cited Soviet geomorphological paper is Markov (1948) and its translated version from 1957. The Soviet literature (Markov, 1948, 1957) reviews all the international

⁶ Although this is not necessarily true as I have shown in the review of the international literature, information which seems that was not available to Sîrcu

⁷ In the same paper, there is a note (page 219) where Martiniuc proposes colines as hills with incipient fragmentation by elongated ridges, and hills as fragmented areas by short hills

⁸ So, they somehow emphasized this term, with which purpose I cannot tell

⁹ Considered as accumulation because in some geological text from before 1930, geological formations from the top of the hills in this area were considered to be "levantine" (the old name of Pontian) lake deposits; this was later refuted

literature but adapt their view to Penck¹⁰ and the regional studies from Russia, introducing the sculptural and structural plains (for the first, the topography cut the geologic structure at a certain angle, while for the second the topography coincides with the geological layer¹¹).

Mihăilescu (1966) predominantly uses the term Jijia Depression but on the map from Figure 29 of his book, splits the depression into three plains: Bahlui Plain, Upper, and Lower Jijia Plains. In Mihăilescu (1977), there are some interesting thresholds for which the author does not always specify the source: the "typical" plains are under 150 or 200 m elevation and have relief under 30 m (low dissected plains) or 75 m (for the heavily dissected plains).

Băcăuanu (1968) geomorphologically studies the region (in his published Ph.D. thesis) and explains the assignation to the plain typology of a region with the relief of *cuestas*, landslides, and hilly geomorphometry, by mentioning that:

- Tufescu (which was the official referent for the thesis) argues for the land use and popular tradition, which Băcăuanu accepts as valid;
- Geomorphologically the plain is a concept that is wider "now" than in the "past", considering the view of the Russian geomorphologists (Markov, 1948, 1957) which name as Russian Plain an area similar to the entire Moldavian Plateau;
- The altitudes are 80% of the surface lower than 200 m which is considered the upper limit of the plains¹², the mean relief is ~70 m, maximum relief values over 200 m are only a few, the ridges have low slopes (1-3‰), the landforms are denudational, the geology is monoclinic and the lithology is clayey-sandy.

Băcăuanu also accepts the usage of Jijia and Bașeu Depression (since Bahlui is already included in the Jijia catchment), which he believes will

supplement the study area characteristics used together with Moldavian Plain. In the Geography of Romania, volume 1 (1983) instead he mentions as a footnote to the Moldavian Plain: "The name of <<plain>> does not refer to the relief, which is generally represented by slightly whitewashed hills, but reflects the characteristics of the geographical landscape, derived from the agricultural function of this region".

Martiniuc and Ungureanu (1970), Coteț (1973)¹³ and Ungureanu (1993) take from Băcăuanu (1968) the term "coline/hilly plain" and consecrates it for some areas of the plateau, a term that we consider nonsense, challenging to digest and which has no argumentation except in the context of the desire to please the pre-and post-communist visions.

2.2 Geomorphometric classifications of landform type

Third, I present the methodology to derive the geomorphometric classification of the Romanian landforms.

At the continental and global scale, the initial landform types that were defined (Penck, 1894; Gregory et al., 1910; Finch and Trewartha, 1942) are: (a) plains, (b) plateaus, (c) tablelands, (d) hills, and (e) mountains. Hammond (1954) mentions four quantitative "attributes" that characterize the land surface and that can be used for landform identification at the continental scale: "local relief, flatness", "characteristic profile", and "characteristic pattern of ridgelines or valley floors". Local relief is the amplitude of elevation in a certain area. Flatness is the "fraction" of a certain "area with inclination less than some chosen boundary value". The characteristic profile is complex, defining the extension from drainage axis to divide, the location along the profile (over or under the median, so in

¹⁰ I was not able to find in Penck's papers the mentions of "Schnittfläche" and "Schichtfläche" for plains; these notions are used in the mountainous areas indeed to separate topography which is purely denudational, from structural influenced topography; probably the soviet authors have translated this approach to plains also

¹¹ From my reading of Penck's papers actually I think that he was referring to the accumulation plains when he was

saying that the topographic surface coincides with the geological strata boundary

¹² This view appears in the Romanian literature without a source; this value appears indeed in Penck (1894) but is not considered a threshold, but rather a statement, that many plains appear under this value

¹³ Although Coteț also has also in paranthesis the Jijia-Balui Depression term and propose the Săveni and Iași Plains as subdivisions

upland or lowland) of the flat land is a good approximation. Characteristic patterns are even more complex, but the average spacing of drainage lines is a good indicator of the texture. Using the first three "attributes", Hammond (1954) proposed a classification in seven categories: nearly flat plains, rolling and irregular plains, plains with widely-spaced hills or mountains, partially dissected tablelands, hills, low mountains, and high mountains. Hammond himself mentions that statistical analysis might be needed to choose boundary values. The calculation of the "attributes" was performed on 13.88 km side squares, which were moved one near each other without overlap. Later, 9.65 km side rectangles were used only for US conterminous (Hammond, 1964a,b). Hammond schema did not use all the possible combinations of the three variables, but a selection was performed using the classes that appeared in the conterminous US. Hammond remarked differences between his results and the literature landform type delineations.

Dikau et al. (1991) is the first raster-based implementation on DEMs of Hammond methodology. This implementation modified some thresholds of Hammond (Table 2) and applied all the 96 possible landform types. The rectangle size was 9.8 km, and the pixel size of 200 m. The validation of the results was performed quantitatively against Hammond maps, differences being remarked.

Gallant et al. (2005) adapted the Dikau et al. (1991) implementation in a focal filter (kernel) approach with square size. The DEM was 1 km in spatial resolution and the kernel size of 10 km. Some modifications of the profile type computation were proposed to resolve inconsistencies that appeared in comparison with the Hammond manual drawn maps.

Karagulle et al. (2017) adapted Dikau et al. (1991) implementation, adding the 50% to 50% percent gentle slope in lowland and upland (Table 3), and removed the plains with other types classes. The circular focal filters were used to compute slope fractions, relief, and profile type fractions but different sizes. The slope was computed using mean values in eight directions with a 3 km kernel, as well as relief. Profile type was computed with a 6 km kernel. The DEM spatial resolution was 250 m. Because the plains were reduced in extension due to

the edge inclusion of the focal filters, a post-processing step was introduced to add over the classification of the plains based on a 5% threshold. Minimum area and noise filtering were applied for the classification and the added plains.

I adopted a circle focal filter on the GMTED 2010 250 m spatial resolution DEM (Danielson and Gesch, 2011) with the 14.5 km original size of the Hammond (1954) and the 9.65 km size used by Hammond (1964a,b). I used a modification of the implementation in GIS proposed by Morgan and Lesh (2005) and Karagulle et al. (2017). I kept the thresholds of Karagulle et al. (2017), which are different than the ones of Hammond (1954, 1964a,b), but similar to the one of Dikau et al. (1991). The classes were modified, their modifications and argumentation being treated further in the discussion section. I did not perform the flat areas optimizations of Karagulle et al. (2017), which requested the modifications of thresholds and kernel windows.

A logical extension of the Hammond approach, also considering its comments (Hammond, 1954, 1962, 1964a,b, 1965) was performed by Iwahashi and Pike (2007) that used slope gradient (in degrees), local convexity (four-neighborhood 3x3 Laplacian filter of DEM grey scaled brightness values that identify the convex areas) and surface texture (a measure of topographic grain, computed as the number of pits and peaks within a radius of 10 cells; pits and peaks are identified by comparing the DEM to a median filter, as the negative and positive differences) to obtain a nested classification using thresholds of mean global values, or in lowland vs. upland (by considering the mean as the boundary between upland and lowland). Their classification is mainly qualitative in naming terrain classes, which mix texture (fine or coarse) with convexity (low or high) and slope (steep or gentle). A mountain will have a steep slope, fine texture, and high convexity, while a plain gentle slope, coarse texture, and low convexity. This classification is different from Hammond's due to the inclusion of texture. It does not consider flatness in the neighborhood, although the computation of the input variables is done in kernel windows (with a size of three pixels for slope, five for concavity, and ten for texture). Since the

classification is unsupervised, the authors did not expect equivalence from traditional landform types. Nonetheless, the density of the classes in a certain area and their arrangement could be used to recognize landform types. The nesting approach allows a certain control of the scale in the sense that on low resolution, it can be applied only to the first level of 8 classes, on the medium resolution the second level of 12 classes, and on higher resolutions, the third level of 16 classes.

3. Results

In Figure 3, the Hammond classification of Karagulle et al. (2017) based on GMTED 250 m data and adaptation to a GIS raster implementation is shown for Romania¹⁴. Our implementation of the Hammond landform classification on GMTED 250 m data is presented in Figure 4 for the 14.5 km focal filter size and in Figure 5 for the 9.6 km focal filter size.

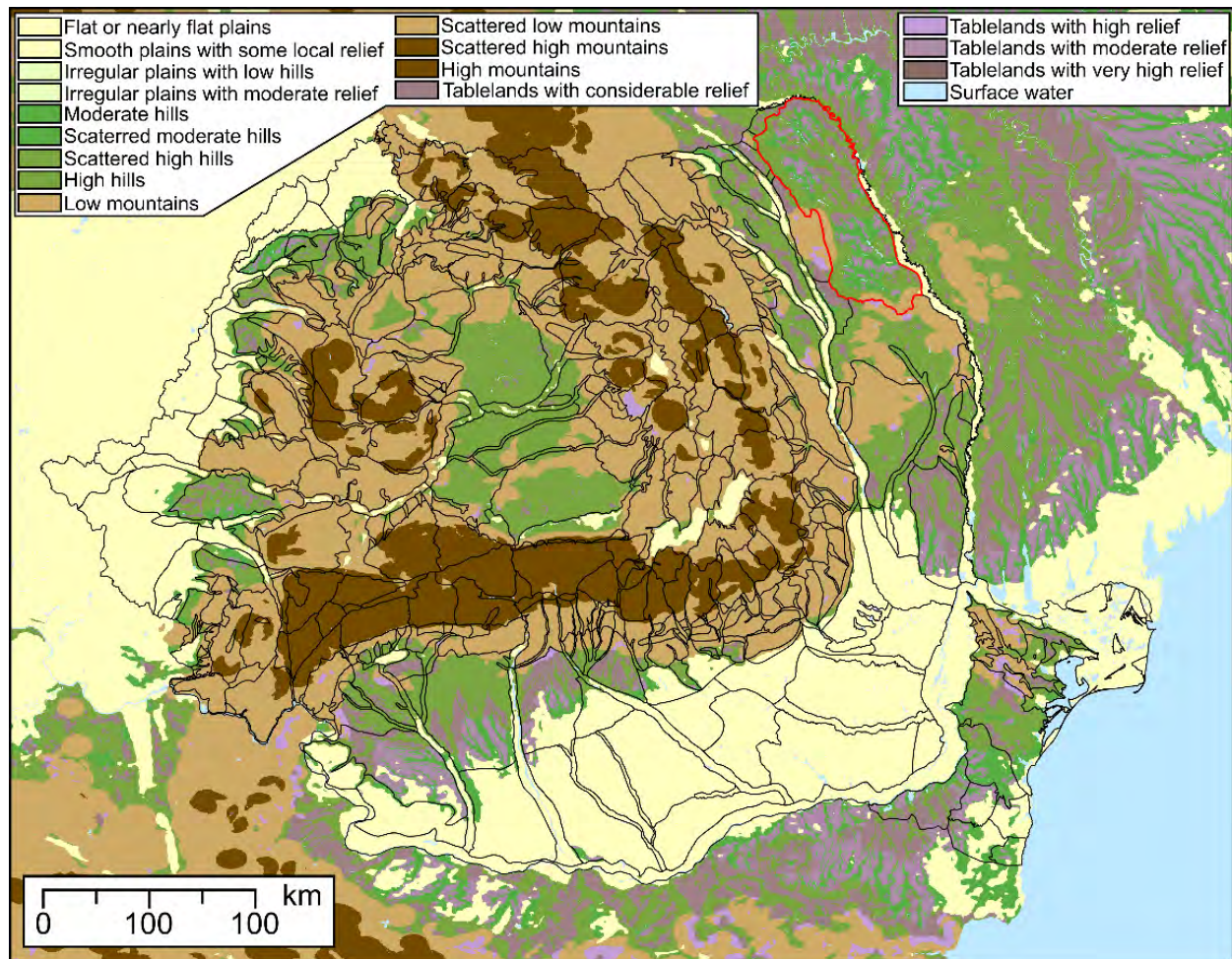


Figure 3 Karagulle et al. (2017) implementation of Hammond's classification for Romania

The Iwahashi and Pike (2007) first level with eight classes is presented in Figure 6. In Figures 3 to 6, over the landform classification was overlain a generalized geomorphological regionalization based on Posea (2006), but retraced in terms of boundaries based on

the 1:200 000 geological maps¹⁵ and the SRTM¹⁶ DEM shading. The idea of a hierarchy of Posea (2006) is welcome in order to be able to distinguish between regions and to be able to apply different criteria, but at present, the criteria are questionable from a

¹⁴

<https://www.arcgis.com/home/item.html?id=cd817a746aa7437cbd72a6d39cdb4559>

¹⁵<http://geo-spatial.org/vechi/download/harta-geologica-a-romaniei-scara-1-200-000>

¹⁶<https://doi.org/10.5066/F7PR7FTF>

hierarchical point of view, requiring a rather geomorphometric methodology, in order to separate the different landforms: plateau, hill, hill, depression, and to merge them in hierarchical units. Otherwise, the hierarchy is subjective, and I have chosen the most used units because the material is also used for students¹⁷.

5. Discussion

Discussing the definitions from the international literature, it can be seen that every author used their own experiences to define the plain: Humboldt (1849, 1856) the field knowledge from several continents that he has visited, by proposing high-altitude plains, Penck (1894) the European continent and Mescherikov (1968) the Russian Plain¹⁸. Even the authors that proposed thresholds of relief, applied

those to their regional knowledge: Penck (1894) to Europe and Finch and Trewartha (1942) to the Fenmann physiographic regionalization. Hammond (1954, 1964a,b) used in his classification relief values under 300 m, but allowed even more relief when used plains with hills or mountains classes. Anyway, all definitions and thresholds are characterized by a certain degree of low slope, flatness, and low relief. Nonetheless, no threshold is based on statistical analysis but rather is inferred from the repartition of regions considered called plains and their appearance on maps. Dikau et al. (1991) are the first that modify the thresholds based on the resolution of the data source (although Hammond was aware of this need), and further modifications are used by the GIS implementation of Gallant et al. (2005), Morgan and Lesh (2005) and Karagulle et al. (2017).

Table 1 Hammond (1954) boundary values for the three terrain attributes; the codes are added for the GIS implementation proposed by Morgan and Lesh (2005)

Flatness			Local relief (m)			Profile		
code	class	% gentle*	code	class	relief	code	class	profile
1000	0	0-10	10	A	0-30	1	a	near all gentle lowland
900	1	10-20	20	B	30-90	2	b	50% gentle lowland and 50% upland
800	2	20-30	30	C	90-300	3	c	>60% gentle lowland
700	3	30-40	40	D	300-900	4	d	>60% gentle upland
600	4	40-50	50	E	>900			
500	5	50-60						
400	6	60-70						
300	7	70-80						
200	8	80-90						
100	9	90-100						

*% gentle slope in the focal filter, the gentle slope being considered the slope <8%

Table 2 Dikau et al. (1991) boundary values for the three terrain attributes; the codes are added for the GIS implementation proposed by Morgan and Lesh (2005)

Flatness			Local relief (m)			Profile		
code	class	% gentle*	code	class	relief	code	class	profile
100	A	>80	10	1	0-30	1	a	>75% gentle lowland
200	B	50-80	20	2	30-90	2	b	50-75% gentle lowland
300	C	20-50	30	3	90-150	3	c	50-75% gentle upland
400	D	<20	40	4	150-300	4	d	>75% gentle upland
			50	5	300-900			
			60	6	>900			

*% gentle slope in the focal filter, the gentle slope being considered the slope <8%

¹⁷

http://www.geomorphologyonline.com/students_materials/GFR/RegionareFizicoGeografica_poligon.kmz

¹⁸ Where the extensive presence of geological platforms coincide with their low altitude and low relief

Table 3 Karagulle et al. (2017) boundary values for the three terrain attributes; the codes are added for the GIS implementation proposed by Morgan and Lesh (2005)

Flatness			Local relief (m)			Profile			profile
code	class	% gentle*	code	class	relief	code	class		
400	0	0-20	10	A	0-30	0	a	<50% gentle slope lowland/upland	
300	1	21-50	20	B	30-90	1	b	>75% gentle lowland	
200	2	51-80	30	C	90-150	2	c	50-75% gentle lowland	
100	3	81-100	40	D	150-300	3	d	50-75% gentle upland	
			50	E	300-900	4	e	>75% gentle upland	
			60	F	>900				

*% gentle slope in the focal filter, the gentle slope being considered the slope <8%

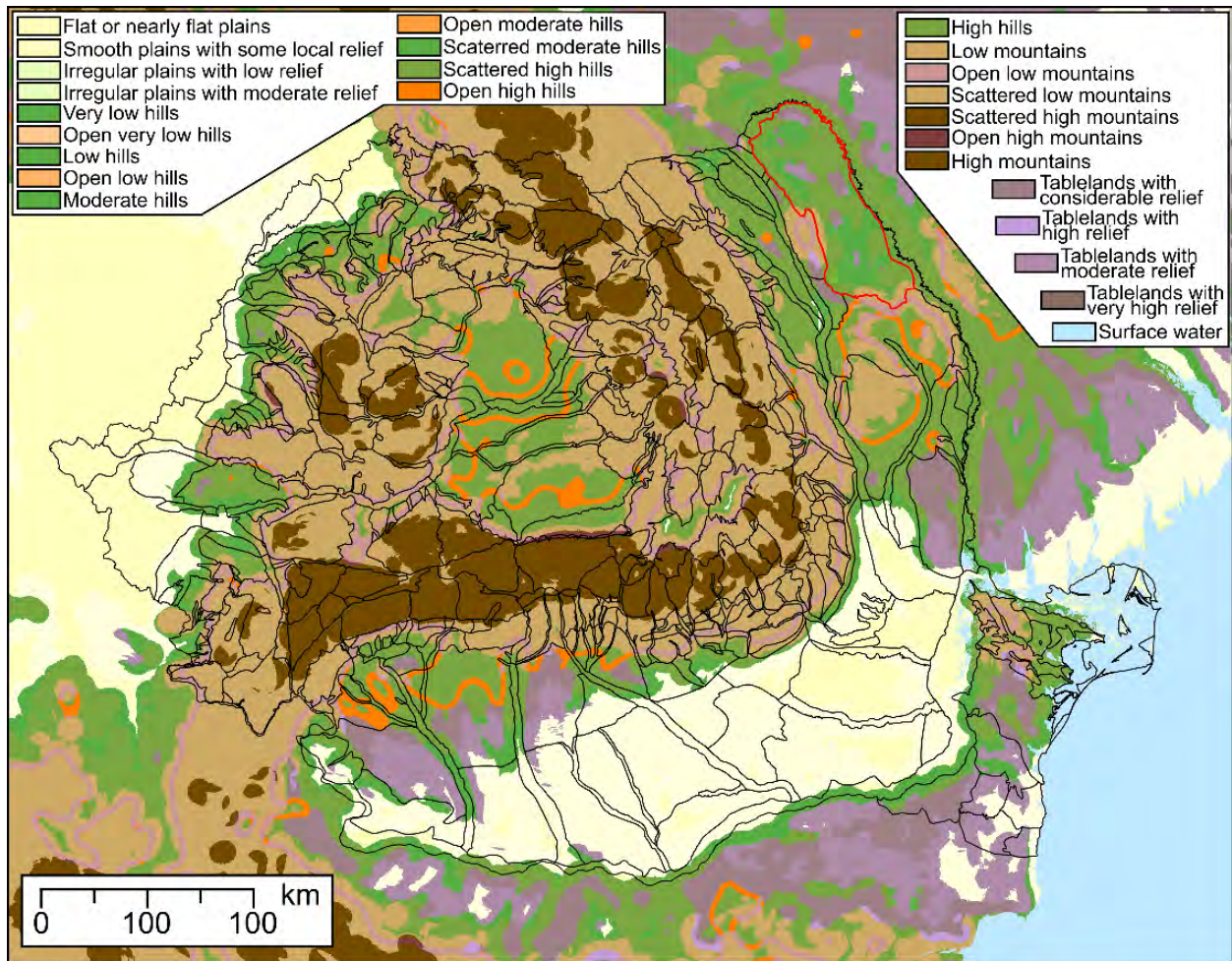


Figure 4 Hammond's (1954) classification for Romania

While Dikau et al. (1991) needed the landform type classification for landslide studies, all the other cited papers were using landform type for ecological classification.

The two world landform type classifications used, with their different implementations, show the difference between the Moldavian Plain's geomorphologic region and other typical plains from

Romania (Fig. 3-6). While this conclusion is very clear, if we look at the maps, some emphasis should be made on several aspects.

First, for regional applications, the thresholds should be adapted (Dikau et al., 1991; Hrvatin and Perko, 2009). There is a great dependence of the slope and relief values on the DEM resolution and focal filter size. Both the variables relate to the limitation of the

rise as the run increase. So, as we use lower resolutions, the slope and gradient decrease, but this reality that was already shown by others (Zhang and Montgomery, 1994; Iwahashi and Pike, 2007) can be resolved by the bigger focal filter sizes, which increase the gradient that can be found in the neighborhood, until the characteristic levels are reached. In Figure 7, I present the variability of relief values for various kernel windows sizes in the case of the study area.

It is clear from this figure that using bigger neighborhood sizes, the majority of the relief is between 100 and 200 m while using smaller

neighborhood sizes, the majority of the relief is under 100 m. From the distribution of relief computed using a 15.25 km diameter circular kernel, the study area's characteristic relief is between 100 and 200 m since relief over 200 m appears as minor frequencies at the border with the higher neighborhood areas.

In Figure 8, I show 15 km swath topographic profiles¹⁹ through the study area on which it can be understood which is the main distribution of relief in the study area.

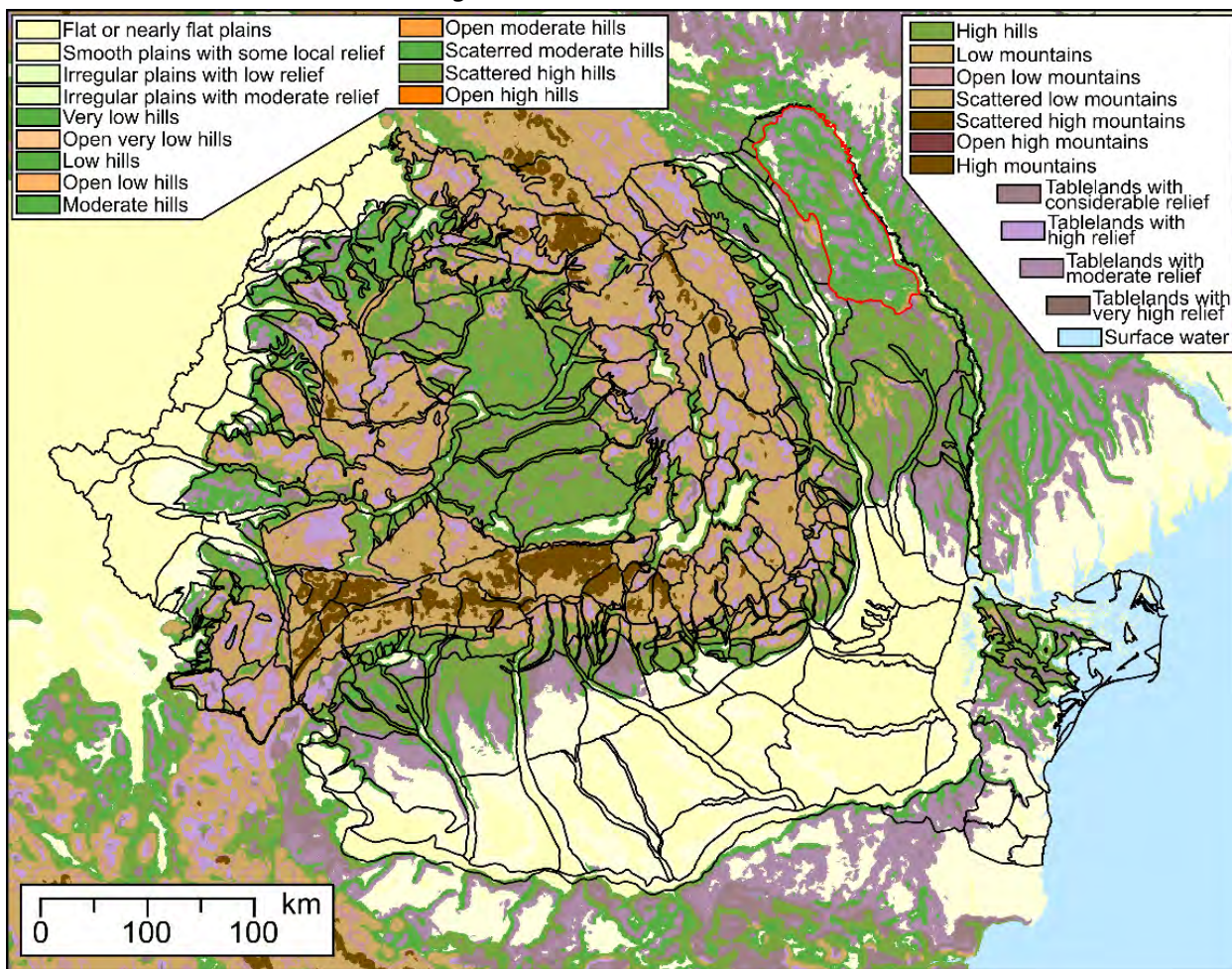


Figure 5 Hammond's (1964a,b) classification for Romania

The 15 km value of the swath width was chosen because this is the width of the interfluvies between the main rivers (Prut, Jijia, Bahlui, Miletin, Sitna). I have also purposely shown the mean values to show

how easily this measure can distort reality. Martiniuc (1955) and Băcăuanu (1968) used mean and maximum values to argue for the low values of relief. Still, the maximum value in the search area

¹⁹ As mentioned by Grohman (2004) the swath profiles are used from the 1920's to derive minimum, mean and

maximum elevations from the surrounding area along a profile path

considered is the definition of relief recognized in the international literature and not the mean value.

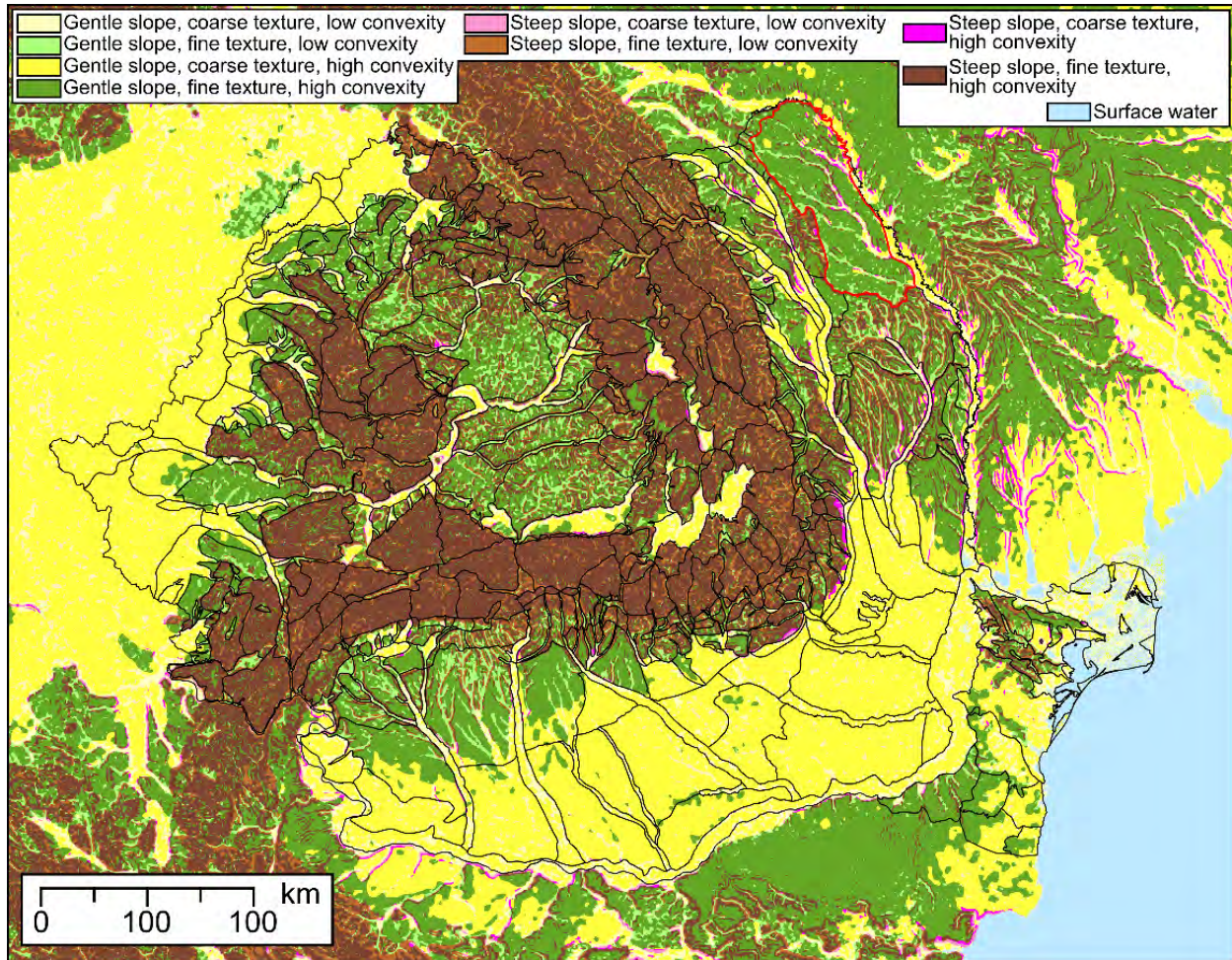


Figure 6 Iwahashi and Pike (2007) classification in eight classes for Romania

Increasing the focal filter size will create unrealistic and spurious features in the classification, especially at borders. Still, it will also find the characteristic gradient values of the area, given by the difference in the regional highest ridges and lowest channels, and not the local gradient as in the case of small focal filter sizes. Using a focal filter reduces the artifacts, but the sensitivity to width remains. Considering the 150 m local relief threshold of Finch and Trewartha (1942), the Moldavian Plain is a plain at 250 m resolution and 3x3 focal filter. Still, it is not at focal sizes higher than 5 km. A 3x3 focal filter will conclude similar to Tufescu's (1935) panoramic view of the Moldavian Plain. The Iwahashi and Pike (2007) classification is not affected by these issues.

Second, the thresholds and the classes proposed by Hammond might be problematic for some areas.

Hammond did not use an objective assignation of the class to the squares used for computation but interpolated the spatial boundaries between classes manually on the map. In the case of objective assignation of the classes by applying the thresholds, as in the case of the raster implementation, some inadvertences appear. In the raster implementations (Fig. 3-5), we can see how all the Subcarpathian areas and some portions of the Someșan Plateau and Moldavian Plateau are classified as low mountains. This is due to the inability of the thresholds to separate high hills from low mountains, especially at bigger kernel size (the low mountains from the enumerated regions are reduced in the 9.8 km kernel size version compared to the 14.65 km one). Also, the plains with moderate relief class occupy considerable areas in the Getic Plateau, Dobrogea Plateau the

Western Hills, where there should be low hills. The plain with hills and mountains class of Hammond that was adapted by Karagulle et al. (2017) as open hills and mountains was a reality only when Hammond interpolated the classes manually. Still, it does not make sense when a raster implementation is used. If a hill or a mountain is rising in a flat area, the 250 m resolution and the ~10 km focal filter size can identify and classify the convexity according to its

morphometry and the flat area as a plain. Only in the situation when there is a parity between flat areas and convexities, or these have small widths, the reality of the class of plain with hills or mountains might be discussed, but, in this case, the flat areas might not be plains but floodplains, and if their width is not big enough the focal filter will remove their influence in the computation of the class.

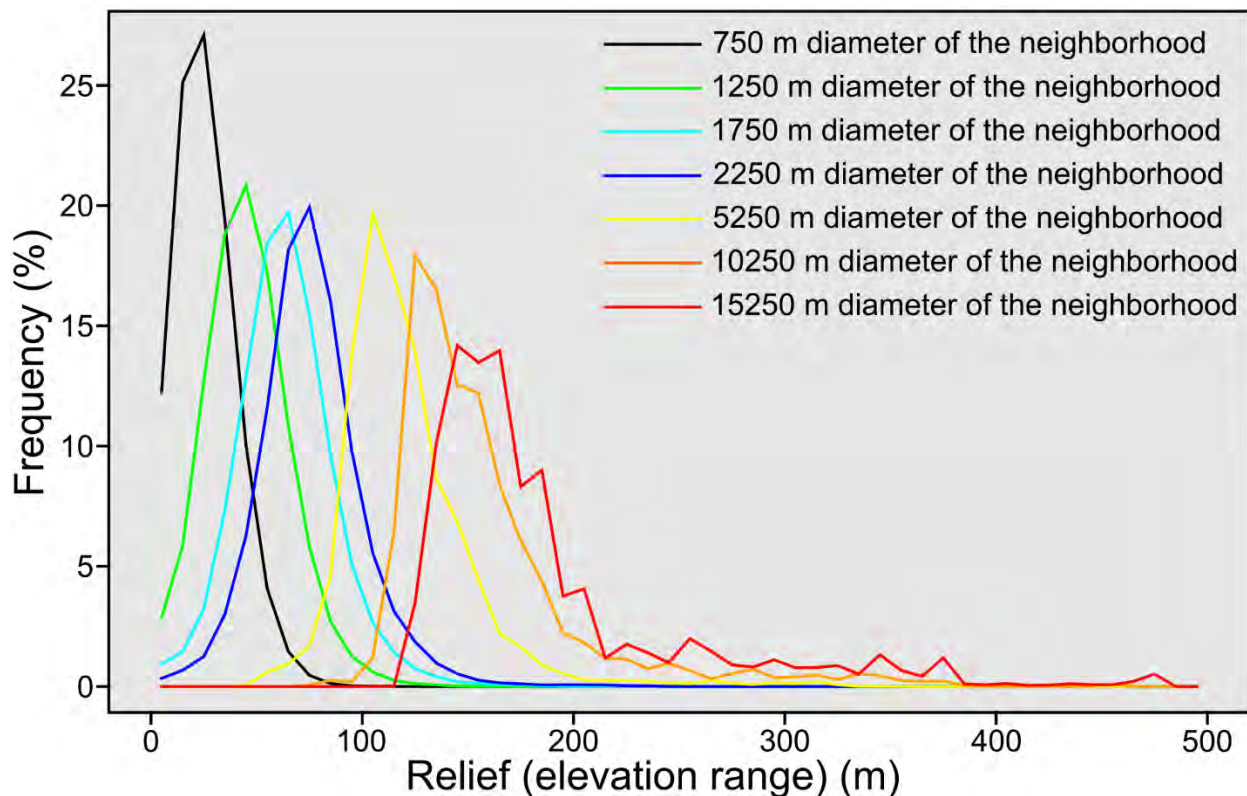


Figure 7 The relief distribution of the study area as a function of the kernel size used for computation

In Figures 3 to 6, it can be seen that the area (the boundary is emphasized with red color) considered as "plain" or "hilly plain" do not meet the geomorphometric conditions of plain, having high slopes over 3° and relief energies that frequently exceed 90 m, and the profile is an upland one. The conclusion of the above is that this geomorphological region should be considered according to its morphometry and geomorphological criteria, the plain attribute being completely wrong. Besides the geomorphometry also the geomorphological processes are not typical for a plain. The cuesta landforms (Ioniță, 2000; Niculiță, 2011), the intense soil erosion, the gullies

and the landslides associated with steep cuesta scarps and other incised hillslopes, the presence of the epigenetic Prut incised floodplain meanders between Rădăuți-Prut and Stâncă-Costești (~60 m of incision) are all geomorphologic aspects not characteristic for a plain. From a geological and palaeogeographical point of view, the area is the morphological result of intense (several hundred meters) erosion of the mudstone facies of the Lower to Middle Miocene sedimentary cover of the Moldavian Platform, being considered by Băcăuanu

1967) a sculptural²⁰ plain, so equivalent to the denudation plains of Fenman (1919) and Finch and Trewartha (1942). The sedimentary cover consists of foredeep deposits folded near the orogen and gently sloped toward the exterior. In the central and northern part of the foredeep basin, the exhumation progressed gradually to the south (from 12 Ma in the north, 10 Ma at Iași latitude, and 7 Ma to the south - de Leeuw et al., 2020) as the basin fill was gently tilted to South-West due to the Northwetsern-Southeastern migrating slab detachment (de Leeuw et al., 2020) that also controls today's subsidence in Focșani Basin (Tărăpoancă et al., 2003). In the last 5.8 Ma since its filling (since Pontian), the area was incised by the river network that was constantly developing toward the south due to the prograding

of the Black Sea coast ~ 100 km (de Leeuw et al., 2020). The timing of the post-Pontian river incision is not established, only the 20-25 m terrace of Bahluiet river at Mădârjești being dated to 20.2 ky BP (Niculiță, 2020). At the border of the study area with the hilly area to the west, there is dated a terrace also on Bahluiet river, related to a fossil landslide that is the oldest date in the Moldavian Plateau: 44-46 ky BP (Niculiță, 2020). Other Pleistocene dates are related to the loess deposits from Mitoc Malu Galben that at 13.5 m thick are dated to ~33 ky BP (Haesaerts et al., 2003).

Regarding the motives for which the geomorphologists accepted and used the term plain for the study area, I argue the influence of political factors based on two reasons.

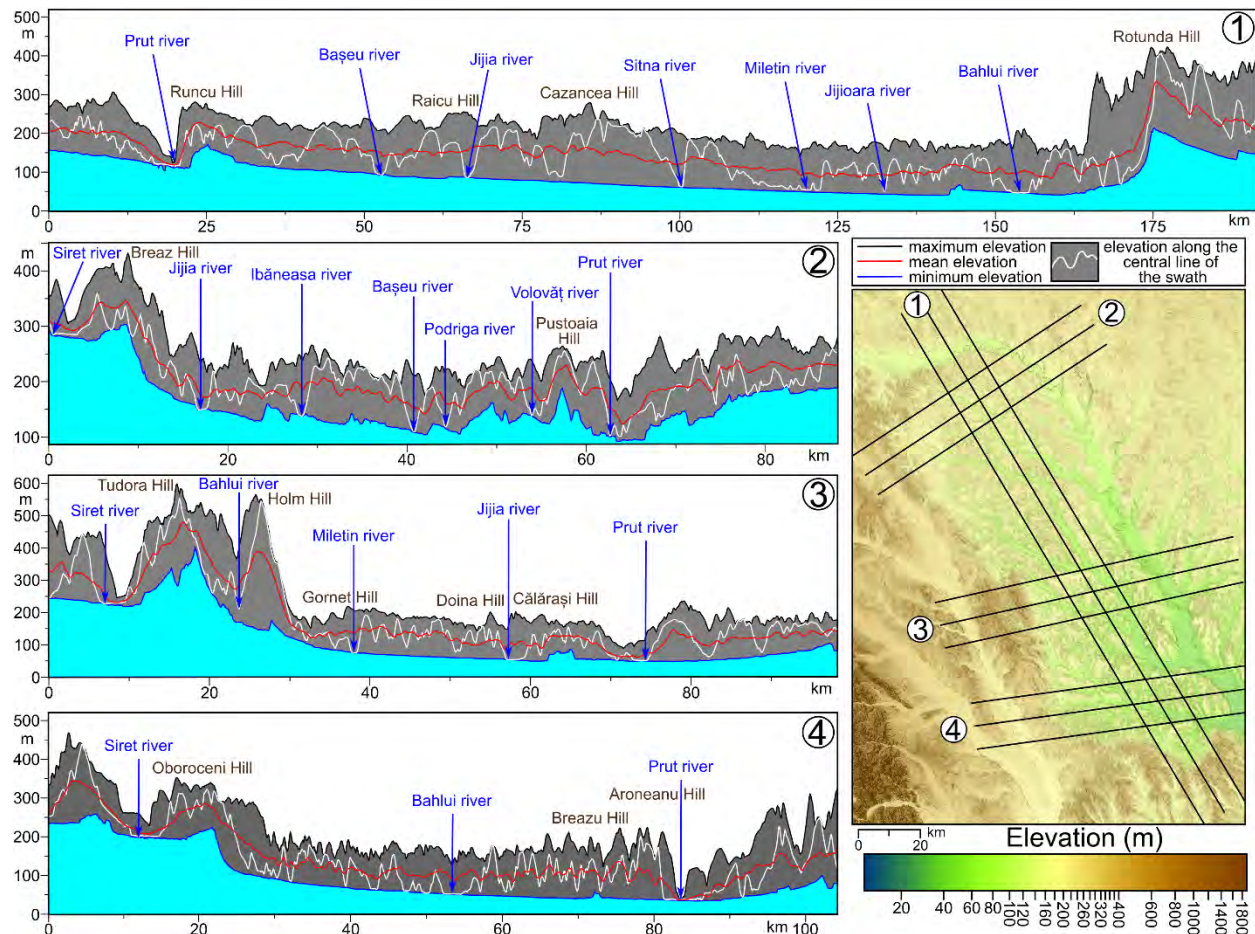


Figure 8 The swath topographic profiles through the study area

²⁰ This term is borrowed from the Soviet geomorphological literature as I have shown below

First, while historically, the first that used the notion of plain is Porucic (1928), the researcher that studied the region and used both terms of depression and plain was Tufescu. He mentions in the 1935 paper on the Jijia Depression boundaries that while previous studies considered the Moldavian Plateau as a region with slight variations in topography (idea argued by Simionescu), actually the area from the Middle Prut River Catchment is lower at the level of the ridges with 150-300 m to the neighboring units. This aspect argues for the consideration of the area as depression or plain. He does not bring further arguments for any of the two categories. Still, the paper keeps the idea of depression as a scientific one and the idea of a plain as a popular one because the locals use the term in connection with agricultural use. But vegetation and land use cannot be used for geomorphologic landform regionalization, even since Penck (1894) being shown that vegetation is not a valid proxy. In 1941, Tufescu comes with the argument that the area should be named as a plain in the geographical regionalization, I think as a way to introduce in the Romanian regionalization popular terms²¹: "It could be said that, compared to the poverty in regional names of our toponymy, the creation of new terms was necessary. And indeed, if for small units the Romanian toponymy is quite rich (and I emphasize that it is infinitely richer than those who drew up our topographic maps, who did not make a purpose out of it), for the name of the larger natural regions, it seems to be quite poor. And I remember in connection with this, the words of a French geographer, who, accustomed to the rich regional nomenclature of France, finds with surprise that there are almost no names of natural regions in Romania. Asking the villagers where they are from, from which region, they only answer by naming their village or

county. << Ainsi la Roumanie est pauvre en noms de lieux, en noms de pays>> (Ficheux, 1929)." (Tufescu, 1941). Further, in the papers, he published after 1945, Tufescu, while using the term depression for Jijia and Bașeu catchments, also use plain as Jijia Plain²² and Moldova Plain²³ (Tufescu, 1966). Suppose we consider the advances of the Romanian language after the Second World War is really not understandable from a geomorphological point of view why to continue using a popular term, the single explanation being the fact that the communist regime pushed for the use of plain in order to argue the need for collectivization of this agricultural area. This is only a hypothesis for which I have no proof at the moment. Another hypothesis is that the geographers tried to please the regime or to increase the study area's agricultural mecanization²⁴. The history, timeline and interpretation of the collectivization can be read in Kligman and Verdery (2011) and Borșa (2013). Tufescu does not bring geomorphologically and geomorphometrically valid arguments for the plain category for the study area, so the conclusion of Sîrcu (1956) of nonsense for the usage of this term I believe is valid.

Second, the note of Băcăuanu in the Romanian Geography volume dedicated to the hilly and plateaus areas (Băcăuanu, 1983) can be seen as an attempt to explain this "anomaly" since in three works regarding the geomorphology of the Moldavian Plain (Băcăuanu, 1958, 1968) and the geography of the Moldavian Plateau (Băcăuanu et al. 1980), despite the relief over 150 m is acknowledged, still, the area is considered a plain. In the case of Băcăuanu's work, considering his geomorphological expertise, I cannot explain the usage of the plain term unless (i) he wanted to please Tufescu, which was an official referent of his thesis, and (ii) the communist party politic was followed, to consider a hilly

²¹ After 1945 Tufescu actually dealt more with human geography than with physical geography, and even before that the geography was seen as a whole so it is not necessarily non-understandable the need to comply with both natural and human criterias; anyway since then this emphasis has changed so there is no need to follow that

²² p. 154

²³ p. 237

²⁴ Here I could mention the personal communication of Grigore Posea during the 30th National Symposium of Geomorphology fieldtrip (Orșova, 29 to 31 May 2014) as the reason for considering plains the Moldavian and Transylvanian low areas, that the number of tractors per surface was assigned based on the landform type; so, for a predominant arable area like the study area the consideration as a hilly area would have been decreased the assigned number of tractors per hectare

agricultural area as a plain to argue for collectivization or (iii) to manipulate the official assignation of tractors based on the landform type (see note 24 on Posea's personal communication). The geomorphologic and geomorphometric arguments that Băcăuanu uses in 1968 can be also refuted since (i) he used for computation of relief 1:100 000 topographic maps and a five by five km square area to compute "maximum" relief as the maximum minus minimum elevation, and further considered as mean relief half of the maximum values, and (ii) used the Markov (1948) views of what is a plain, ignoring the rest of the international literature. The genesis of an area (denudational) and its lithology cannot be used as a criterion since plains are various in genesis, structure, and lithology (Markov, 1948). Here I would add that also the Soviet supervision could have been played a role in the decision of using and arguing for the term plain, considering the note of Cotet and Martiniuc (1960) that L. G. Kamanin informed them verbally in the field about the typology of coline plains that can/should(?) be attributed to area from the Moldavian Plateau.

6. Conclusions

The main conclusion is that in the case of the geomorphologic region in the discussion, called Moldavian Plain (Băcăuanu, 1968, 1983), the attribution to plain should be removed, at least from the geomorphologic regionalization. I propose the use of the geomorphologic terminology from before 1945: Jijia Depression or Jijia Hills. Restitution refers to the acknowledgment that political/subjective decisions influenced the decision regarding the typology and refuted all geomorphologic arguments.

Acknowledgments

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poster, which also remarked the "hilly plain" problem, pointing me to understand the need for restitution. The comments of two anonymous reviewers are acknowledged since they triggered a wider search of the literature and a wider consideration of motives for the geomorphologic anomaly. Thanks to Igor Florinsky for the help with Russian literature.

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Évaluation de la susceptibilité des sols au ruisseau dans le bassin versant du Knam (Ouest Cameroun), en appliquant la méthode Services de Conservation des Sols Numéro de Courbe

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RÉSUMÉ

Le ruissellement pluvial en zone de montagne constitue un problème environnemental de plus en plus préoccupant. Dans le bassin versant du Nkam où on observe une forte dynamique paysagère, Il est à l'origine de phénomènes violents tels que l'érosion des sols, les inondations ou les coulées boueuses. Il réduit de plus en plus les terres cultivables et les capacités de réservoirs d'eau et, affaiblit les revenus des populations. Partant de ce constat, le présent article évalue la susceptibilité des sols de ce bassin au ruissellement. La méthode d'évaluation développée associe le Système d'Information Géographique (SIG) à la méthode Services de Conservation des Sols Numéro de Courbe (SCS CN), basé sur les caractéristiques physiques, l'utilisation, et les conditions hydrologiques des sols sont utilisés. À cet effet, les cartes des sols, des pentes et de l'occupation du sol ont été réalisées et croisées pour déterminer le niveau de sensibilité des sols au ruissellement. Les résultats obtenus indiquent que 58% du bassin est constitué de pentes > 15%, les sols y sont constitués de minéraux facilement altérables. Une régression forestière de l'ordre de -2,79 % par an entre 1978 et 2019 a été constatée. La combinaison de ces trois facteurs indique que les sols de la région étudiée ont atteint un seuil inquiétant d'exposition au ruissellement (CN = 67). Le reboisement et les techniques de culture sans labour en amont du bassin atténueraient l'effet de ruissellement.

MOTS CLÉS Ruissellement, sol, érosion, SCS CN, bassin versant du Nkam, Ouest Cameroun

ABSTRACT

Runoff in mountain areas is an increasingly worrying environmental problem. In the Nkam watershed where we observe a greater change in the landscape, it is at the origin of certain phenomena such as soil erosion, floods and

mud flow. It increasingly reduces arable land, decreases the capacity of water reservoirs and weakens the income of the populations. Based on this observation, this article assesses the susceptibility of the soils in this basin to runoff. The evaluation method developed to carry out this investigation combines the Geographic Information System (GIS) with the Soil Conservation Services Curve Number (SCSCN) method, based on the physical characteristics, land use and hydrological conditions of the soils used. With these, the soil, slope and land use maps were produced and superposed to determine how sensitive the soils are to runoff. The results obtained indicate that 58% of the basin is made up of slopes > 15%, the soils are made up of easily weathered minerals and the rate of forest regression is - 2.79% per year between 1978 and 2019. The combination of these three factors indicates that soils in the study area have reached a worrying threshold of exposure to runoff (CN = 67). Reforestation and no-till cultivation techniques upstream of the basin would mitigate the runoff effect.

KEYWORDS runoff, soil, erosion, SCS CN, Nkam watershed, West Cameroon

1. Introduction

Le sol est le siège des processus biologiques vitaux pour la continuité de la vie sur Terre. Il constitue, en effet, le fondement de la quasi-totalité de notre alimentation et de nos ressources forestières. Cependant, depuis plus de 40 ans, 30% de terre dans le monde sont devenues non-productives (Pimentel, 2006) et cette non-productivité a été attribuée à l'érosion des sols par ruissellement (El Hage Hassan, 2013). Ce phénomène hydrologique naturel (ruissellement) qui parfois prend des allures d'aléa naturel, est favorisé par les pratiques culturelles de déforestation et d'agriculture non adaptées aux conditions physiographiques locales (pente, sol et climat) (Djoukeng et al., 2015). Ceci est d'autant plus vrai pour les bassins versants tropicaux intramontagnards à sols fragiles soumis à des précipitations intenses. Au niveau du bassin versant du Nkam, comme dans toute la région des Hautes Terres de l'Ouest Cameroun (HTOC), l'agriculture reste pour les populations, la principale activité économique (Ngoufo, 1992; Ngoufo et al., 2001; Kaffo, 2005; Kaffo et Fongang, 2009; Lounang et al., 2014). Cette agriculture reste largement extensive et l'attachement des populations aux pratiques traditionnelles constituent des obstacles à l'amélioration de la fertilité des sols. Il s'en suit donc une dégradation continue des sols par ruissellement qui limite le développement agricole du bassin. Cependant, l'appréhension du phénomène de ruissellement reste encore incomplète bien qu'elle ait déjà fait l'objet de nombreuses études (Sadiki et al., 2004; Kacem et al., 2014; Kacem et al., 2017). Cela vient probablement du fait que les paramètres

explicatifs de ce phénomène diffèrent d'un lieu à l'autre et leur estimation reste difficile à l'échelle du bassin versant. L'évaluation de la susceptibilité des sols au ruissellement dans le bassin versant du Nkam est donc l'objectif de cet article qui nécessite l'appréhension des caractéristiques de la surface du sol (pente, type de sol, utilisation du sol, etc.). Afin de mieux modéliser la susceptibilité des sols au ruissellement, la présente étude utilise la méthode SCS CN associée aux SIG dans le but de cerner les prédispositions naturelles et anthropiques qui caractérisent l'exposition des sols. Il s'agit d'une approche cartographique qui intègre et fait interagir les facteurs qui influencent le ruissellement dans les régions de montagne. C'est notamment le cas du bassin versant du Nkam située sur les HTOC. L'hypothèse qui sous-tend cette recherche postule que l'exposition des sols au ruissellement dans le bassin versant du Nkam est tributaire à une forte régression du couvert végétal induite par l'action humaine sur les terrains pentus aux sols facilement altérables. Ainsi, à la suite d'une présentation rapide de la zone d'étude, on procédera à la cartographie de la vulnérabilité des sols au ruissellement suivie d'une discussion des analyses et de l'interprétation des résultats.

2. Matériels et Méthodes

2.1. Localisation géographique

Située sur les HTOC entre le 5°01'03.55" et 5°38'08.08" de latitude nord et, entre 9°45'44.57" et 10°21'12.95" de longitude Est, le bassin versant du Nkam draine une superficie d'environ 2 348 km² soit environ 19,4% de la superficie totale du bassin du

Wouri (11 700 km²). Ainsi, il draine les Départements de la Menoua et du Haut Nkam dans la Région de l'Ouest et le Mungo dans celle du littoral (Fig. 1). Sur le plan de la géomorphologie, il s'étend sur trois grands domaines géomorphologiques d'amont en aval. La partie amont du bassin, de 1 700 m à 2 681 m d'altitude, est celui du mont Bambouto qui est un massif volcanique aux pentes fortes (Déruelle et al., 1991). La partie centrale et orientale du bassin versant est une zone de plateaux et fait partie du plateau Bamiléké située entre 1 400 et 1 600 m d'altitude. La partie aval du bassin est une zone de plaine aux pentes faibles et au relief plat. Avec une altitude variant de 449 m à 2 681 m. Géologiquement, les trachytes et les basaltes sont les formations les plus répandues (Momoa et al., 2016). La lecture des données climatiques de la station de Dschang, indique une température moyenne annuelle de 20,4 °C. Mars est le mois le plus chaud avec 21,8 °C. Son régime pluviométrique est monomodal avec des pluies torrentielles d'août à octobre. Des intensités dépassant les 150 mm en 24

heures ont été observées (Nanfack, 2021) provoquant des crues violentes, ayant de répercussions directes, tant à l'amont par des pertes de terres arables (Djoukeng et al., 2015) qu'à l'aval par les inondations, les coulées de boue et les dépôts de sédiments (Nanfack et Julius, 2020). Le réseau hydrographique est particulièrement dense. Avec environ 620 800 habitants 3^e RGPH (2010) et une densité moyenne d'environ 264 hab/km², le bassin versant du Nkam reste très convoité. Les conditions écologiques particulièrement favorables au développement de l'agriculture expliquent que les noyaux de densités les plus élevées correspondent aux hautes terres volcaniques du bassin versant (Dongmo, 1981). Ces espaces sont de plus en plus exploités à des fins agricoles, minières ou encore urbaines. Cette anthropisation croissante s'accompagne d'une fragilisation des équilibres naturels caractérisée par une régression constante des espaces forestiers qui à son tour expose les sols à de vaste campagne d'érosion hydrique des sols.

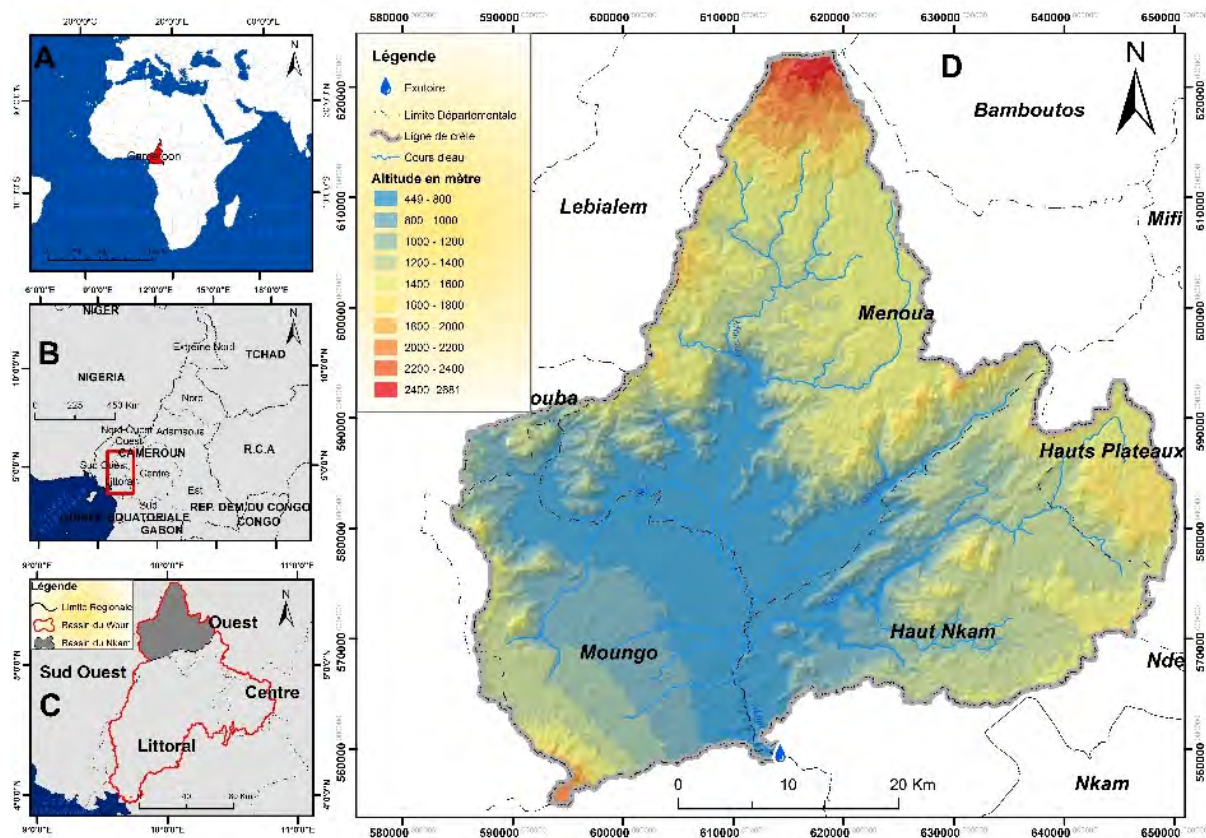


Figure 1 Localisation et présentation de la zone d'étude

2.2 Données et matériels

Le problème de ruissellement des eaux pluviales pose des dégâts économiques et écologiques et même sociaux dans le bassin versant du Nkam. La démarche appliquée dans cette étude pour cartographier les zones susceptibles au ruissellement a nécessité une compilation de données cartographiques (l'image radar SRTM, images satellite Landsat MSS, ETM+ et SPOT5 et la carte pédologique) de la zone étudiée. La collecte de ces informations a permis de structurer une Base de Données à Référence Spatiale. Les logiciels ArcGIS 10.2 et Envi 5.3 ont été utilisés pour le traitement des données.

2.3. Méthode

Pour élaborer la carte de susceptibilité des sols au ruissellement, la méthodologie développée par le Service de Conservation des Sols américain (SCS) qui se base sur les caractéristiques physiques, l'utilisation du sol, et les conditions hydrologiques des sols, a été mise à contribution. Selon cette méthode, les caractéristiques du bassin sont intégrées et quantifiées selon une échelle. Les numéros se retrouvant sur cette échelle sont appelés numéros de courbe et sont en relation avec la capacité d'infiltration (Smith et Eggert, 1978). La maxime sur laquelle elle repose postule que le ruissellement augmente proportionnellement à l'accroissement du numéro de courbe (CN). Cette méthodologie comprend les étapes suivantes présenté ci-dessous.

2.3.1 Classification des sols selon les grands groupes (A, B, C, D)

La méthode SCS stipule que les sols doivent être classifiés en fonction de leur capacité d'infiltration. Cette capacité étant inversement proportionnelle à la propension au ruissellement, cette classification permet de connaître le potentiel que présente chaque sol face au ruissellement (Gassman et al., 2007). Dans le cadre de ce travail, la carte pédologique du bassin versant du Nkam issue de la Cartothèque de l'IRD est utilisée. Quatre grandes classes de sol sont ainsi créées pour le bassin du Nkam en se basant principalement sur la texture. Les sols ferrallitiques qui bénéficient d'un bon drainage

interne forment la classe A; leur capacité d'infiltration est très élevée. Les sols riches en humus et les sols peu évolués noirs sur cendre présentent généralement un taux de percolation et d'infiltration moindre, ils se retrouvent en classe B. La classe C est constituée des sols moyennement organiques associés aux sables fins mal drainés, de sols limoneux et de sols perméables mais minces. Les sols hydromorphes constitués des argiles lourdes mal structurées et mal drainées se retrouvent en classe D.

2.3.2. Utilisation du sol

On note diverses utilisations des sols dans le bassin versant du Nkam. La propension au ruissellement augmente de la forêt aux surfaces artificialisées en passant par les surfaces agricoles. La classification est basée en bonne partie sur l'état de la couverture végétale. C'est dire que, les surfaces offrant une meilleure couverture au sol telles que les forêts présentent une tendance au ruissellement plus faible. Par exemple, un champ laissé en jachère contrairement à une surface forestière, présentera une couverture végétale peu dense et l'interception y sera peu importante ce qui ne favorise guère l'infiltration. A l'aide de l'image satellite SPOT5 de 2019, les différentes unités d'occupation du sol ont été réalisées pour le bassin versant du Nkam.

2.3.3. Importance de la pente

La pente joue également un rôle important dans la méthode SCS Curve Number pour l'évaluation du ruissellement. Cependant, cette méthode fut développée pour des pentes variant de 3 à 8 % (Tramblay, 2012) ; alors que la gamme rencontrée dans le bassin versant du Nkam est beaucoup plus étendue. Les sols de la plaine des Mbo ont en effet une pente bien inférieure à 3% et certains sols des zones montagneuses présentent des valeurs supérieures à 8%. Les sols à faible pente sont donc considérés comme des sols facilitant l'infiltration. Pour les sols ayant une pente supérieure à 8%, les pires conditions sont utilisées.

2.3.4. Conditions hydrauliques

Un autre critère s'ajoute aux précédents, celui de la condition hydrologique qui varie de pauvre à bonne. Cette condition qualifie l'état de la couverture

végétale pour une même utilisation du sol. Un couvert bien établi qui restreint le ruissellement sera associé à une bonne condition hydrologique. L'absence du couvert végétal pour les surfaces artificialisées et agricoles aura pour effet de détériorer la condition hydrologique.

L'utilisation du sol et la condition hydrologique font aussi le sujet de quelques hypothèses (Garen and Moore, 2005). La méthode SCS Curve Number définit des valeurs de CN pour un très grand nombre de cultures et autres utilisations du sol. La variation entre deux unités d'occupation du sol est bien souvent moindre que la variation due à la condition hydrologique. Comme l'évaluation de cette dernière est subjective, il devient alors peu significatif de

détailler l'utilisation au niveau de chaque unité d'occupation du sol. Des regroupements ont donc été faits : la forêt et la savane ont été réunies et sont traitées comme des surfaces boisées. Elles sont caractérisées par une bonne couverture végétale. Les terres cultivées et les terres laissées en jachère ont été réunies en surface agricoles. Le couvert végétal y est peu dense. Le bâti, les routes bitumées et les sols nus ont été regroupés en surfaces artificialisées. Elles sont caractérisées par une absence du couvert végétale. Le Tableau 1 résume les différentes valeurs de CN suite aux regroupements et hypothèses effectués. Ces valeurs sont extraites des résultats fournis par le Service de Conservation des Sols des USA et certaines interpolations ont été réalisées.

Tableau 1 Numéro de courbe (CN) pour les différentes combinaisons sol, utilisation du sol, pente et condition hydraulique

Utilisation du sol	Pentes (%)	Conditions hydrologiques	Classe de sol			
			A	B	C	D
Surfaces agricoles	<3%	Pauvre	63	74	81	82
		Bonne	60	70	78	81
	3-8%	Pauvre	65	76	84	88
		Bonne	63	75	83	87
	>8%	Pauvre	72	81	88	91
		Bonne	67	78	85	89
Végétation dense	<3%	Pauvre	25	55	70	77
		Bonne	22	53	65	74
	3-8%	Pauvre	41	63	75	81
		Bonne	25	55	70	77
	>8%	Pauvre	47	68	80	84
		Bonne	41	63	75	81
Surfaces artificialisées	Surfaces artificialisées denses		73	83	88	90
	Surfaces artificialisées peu denses		59	74	82	86

Source : Service de Conservation des Sols du ministère de l'Agriculture des États-Unis interpolé par les auteurs.

2.3.5. Évaluation de la superficie de chaque combinaison

L'évaluation de la superficie de chaque sous-groupe, soit chaque combinaison classe de sol et utilisation du sol constitue la cinquième étape ; les valeurs en relation avec l'utilisation du sol, la pente, les conditions hydrologiques et les types de sol ont été introduites dans un logiciel de SIG (ArcGIS version 10.2) pour l'évaluation de la superficie de chaque combinaison classe de sol et utilisation du sol. La dernière étape est celle de la compilation du coefficient de ruissellement (CN) pour tout le bassin versant du Nkam en calculant une moyenne

pondérée selon la superficie de chaque combinaison de sol (A, B, C et D) et d'utilisation du sol (boisée, agricole et artificialisée) suivant la relation ci-dessous:

$$CN_{moyen} = \frac{\sum_{i=1}^n A_i CN_i}{\sum_{i=1}^n A_i} \quad (1)$$

où, CN moyen = coefficient de ruissellement moyen; i = un type et une couverture du sol (forêt, agricole, urbain); A_i = une surface de drainage correspondant à la surface occupée par i; CN_i = un coefficient de ruissellement CN correspondant à A_i

Le résumé de cette méthodologie est présenté à travers la Figure 2 ci-dessous.

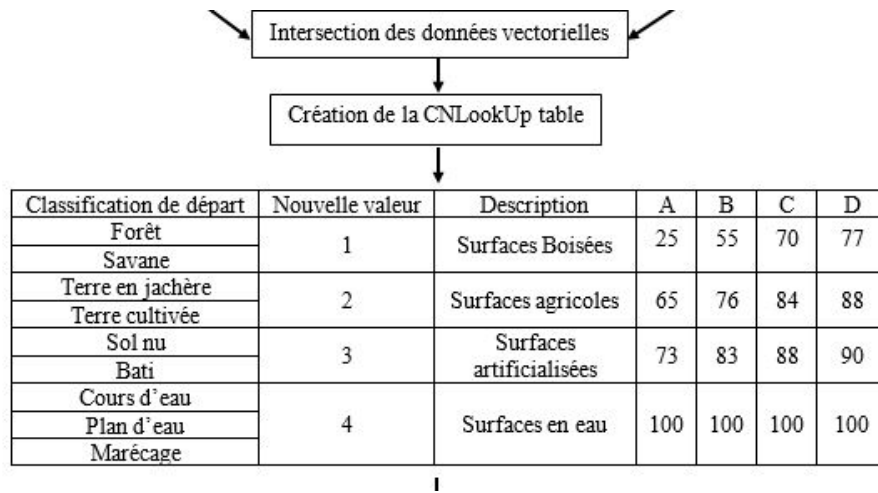


Figure 2 Méthodologie de réalisation de la carte de susceptibilité des sols au ruissellement

3. Résultats

3.1. Les facteurs de prédisposition des sols au ruissellement

3.1.1. Des sols aux minéraux facilement altérables

La donnée pédologique est une variable de contrôle du phénomène de ruissellement. Le bassin versant du

Nkam est constitué d'une mosaïque de sols liés à l'histoire géologique de la région des HTOC. Ils sont issus de basaltes, de terrains sableux et d'alluvions (Fig. 3). Les sols représentés sur cette Figure 3 peuvent être regroupés en quatre types de sol. Les sols peu évolués qui couvrent une faible superficie.

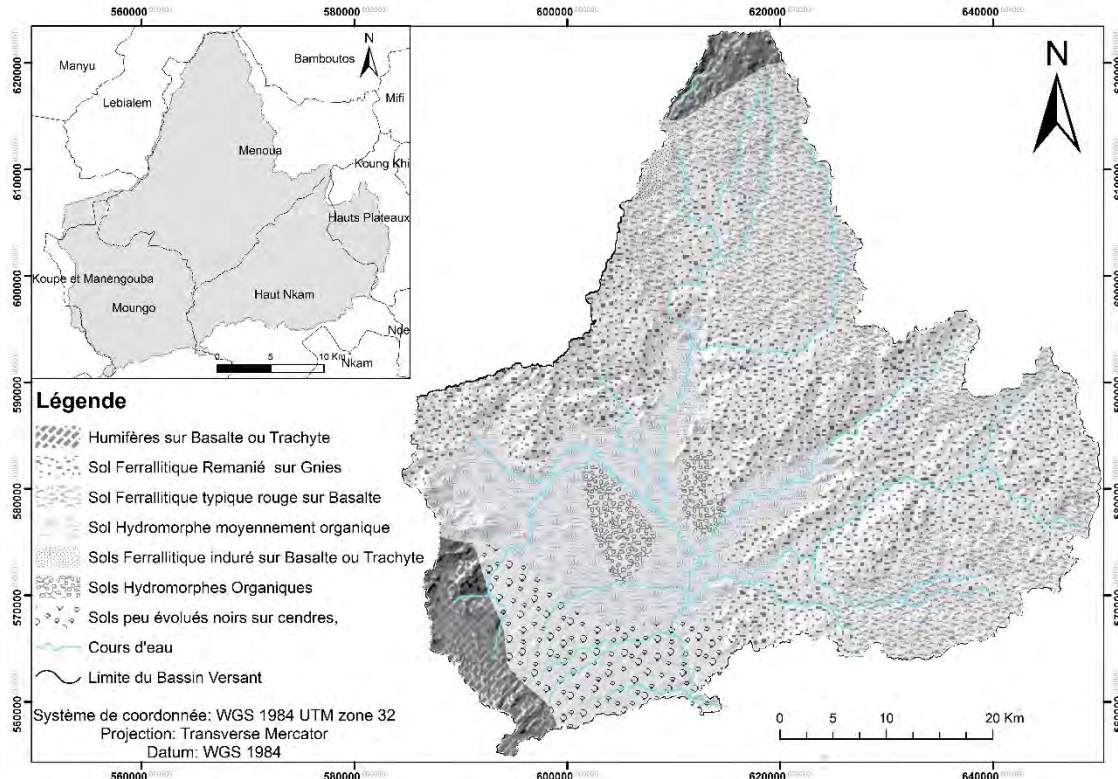


Figure 3 Pédologie du bassin versant du Nkam

Les profils sont peu épais et on trouve de nombreux blocs de basalte bulleux en surface et dans le profil. Ces sols sont désavantagés par leur faible profondeur. Les sols ferrallitiques sur basalte quant à eux sont particulièrement bien représentés à des altitudes supérieures à 700 m. Ces sols sont caractérisés par une texture très argileuse (70 à 80% d'argile) qui assure une bonne capacité de rétention d'eau et en même temps une perméabilité suffisante. Ces sols sont formés sur des roches volcaniques qui affleurent à mi-pente du Manengouba et du plateau bamiléké. Les sols humifères sur basalte ou trachyte y sont peu cultivés et se trouvent le plus souvent sous savane ou sous forêt hétérogène. Ils occupent de longues collines régulières. La plaine par contre est recouverte de couches alluvionnaires. Ces dépôts fertiles sont liés à la dynamique fluviale et au fonctionnement du Nkam, et sont à l'origine des activités anthropiques dominantes dans la plaine comme le maraîchage.

3.1.2. Une pente favorable aux ruissellement

Les conditions d'infiltration de l'eau dans les aquifères tiennent compte de la pente. C'est l'un des facteurs à l'origine du ruissellement. Elle influence le transfert et l'accumulation du ruissellement mais pas de la même manière. Les pentes fortes sont favorables au transfert rapide des eaux, tandis que les pentes faibles sont favorables à l'accumulation des dépôts charriés par les eaux de ruissellement. La distribution spatiale des pentes dans le bassin versant du Nkam est présentée à la Figure 4.

La pente influence directement le débit de pointe. C'est un élément déterminant dans l'orientation et l'organisation des cours d'eau. Cette Figure 4 a permis de mettre en évidence la variation de cette dernière dans le bassin du Nkam. Dans ce relief de montagne, environs 58% des terrains sont sur des pentes supérieures à 15% contre seulement 42% de terrain sur pentes inférieures 15%.

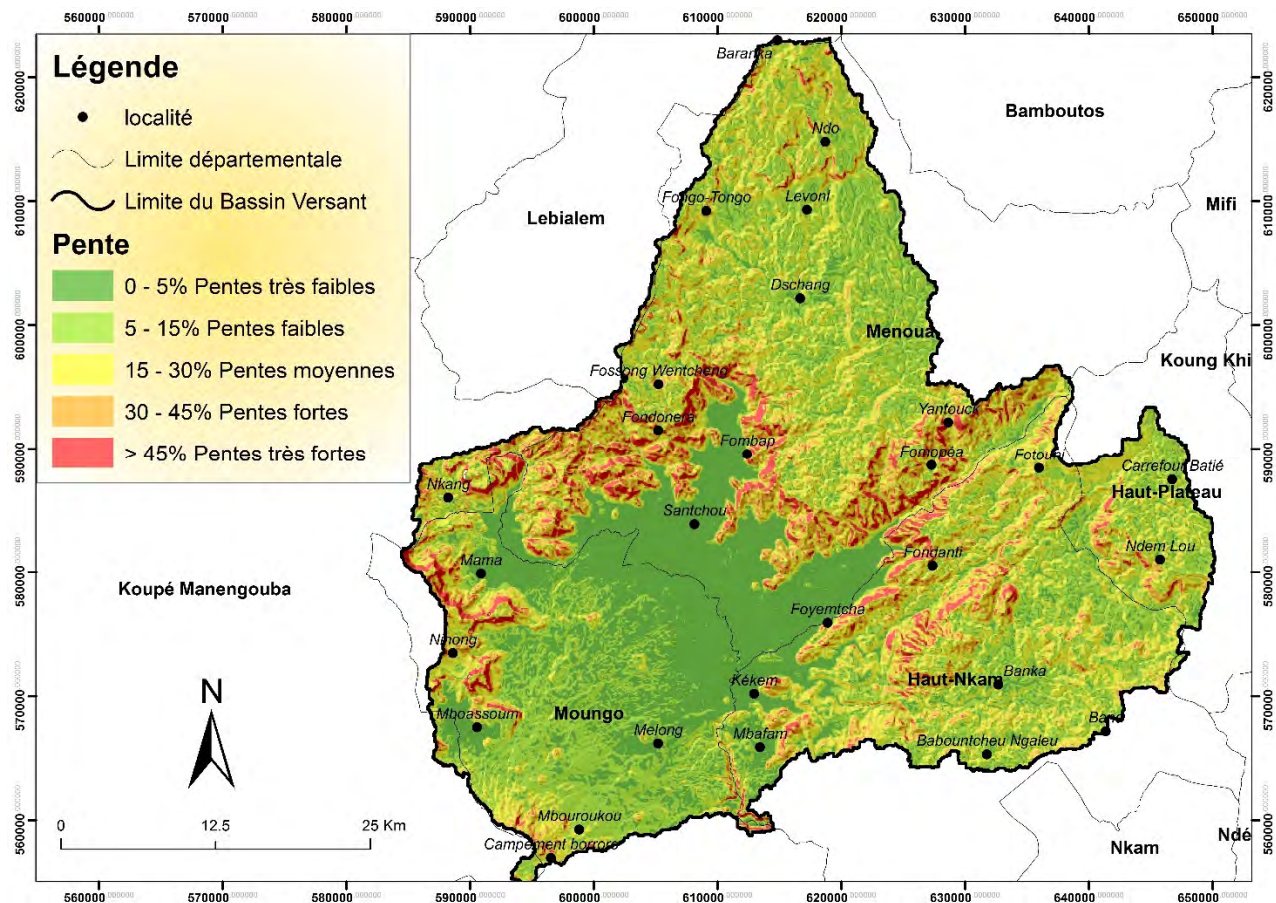


Figure 4 Carte des pentes du bassin versant du Nkam

La structure morphologique du bassin oblige souvent les affluents du Nkam à dévaler les fortes pentes et en même temps à stagner dans les zones à très faibles pentes (Plaine de Mbo) pour créer des inondations. La pente moyenne vaut 12,93% et l'indice de pente global, 5,36 m/km.

3.1.3 Une régression considérable des surfaces forestières

L'étude diachronique des unités d'occupation du sol est captivante dans ce travail, car elle révèle le niveau de sensibilité actuelle du bassin versant. La forte artificialisation des espaces, réalisée de manière anarchique, aggrave les conséquences des événements naturels et particulièrement, le ruissellement. Avec la cartographie du bassin du Nkam, l'analyse diachronique de l'occupation du sol de 1978 à 2015 a été réalisée suivant la méthode de comparaisons post-classificatoires. L'analyse spatiale a permis de mesurer les états successifs de l'occupation du sol et de décrypter les fluctuations

des éléments de cette occupation depuis 1978 (Fig. 5).

On peut admettre, à travers la Figure 5 que la forêt était l'unité d'occupation du sol la plus rependue dans le bassin en 1978. Jusqu'à cette date, l'étendue de forêt ne semblait pas avoir été sensiblement modifiée par les habitants qui étaient peu nombreux et qui vivaient en quelque sorte en harmonie avec les paysages naturels. Au cours des années 80, l'augmentation importante de la population¹ dans le bassin a entraîné une réduction des surfaces forestières. Cette régression n'a pas cessé de se poursuivre et de s'accroître pour atteindre un rythme de déforestation de -2,78% ; soit une perte considérable d'environ 20 km² de forêt par an pour la période 1978-2019. La Figure 6 présente cette régression spectaculaire des forêts au profit des surfaces artificialisées et agricoles en seulement 40 ans.

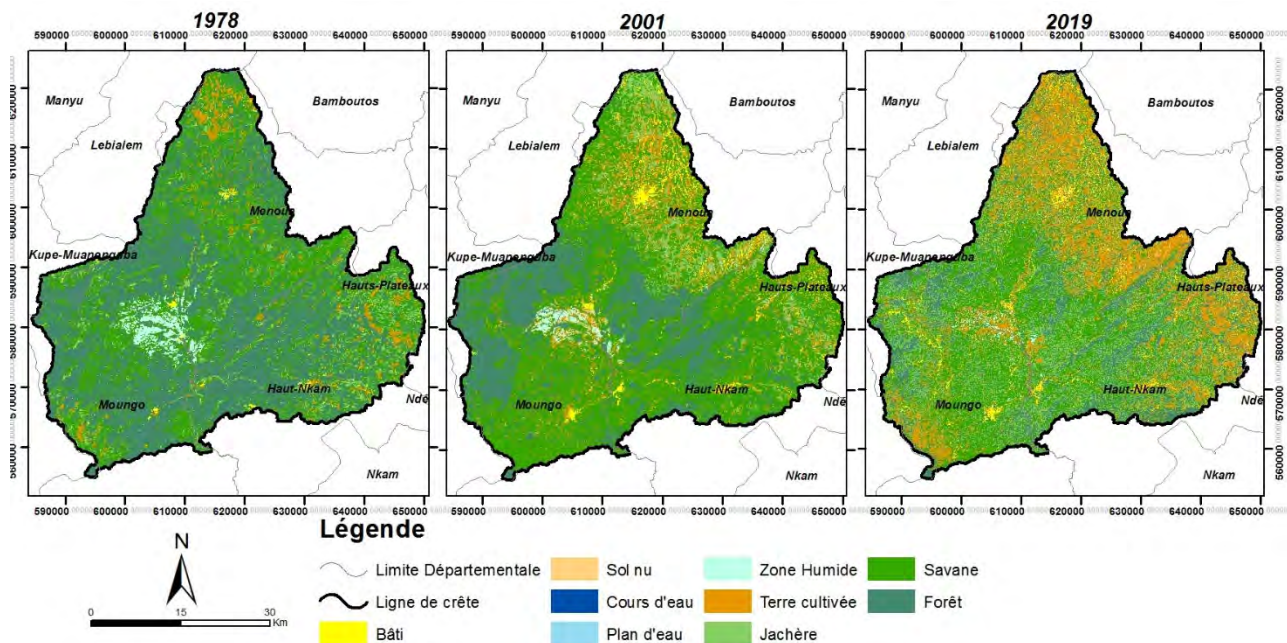


Figure 5 Evolution de l'occupation du sol dans le bassin versant du Nkam

L'intervention de l'homme en tant qu'agent de modification de l'environnement du bassin du Nkam est donc multiforme au regard de cette Figure 6. La dégradation des forêts par celui-ci entraîne de profondes perturbations dans le régime

hydrologique du bassin (torrentialité, inondation, envasement etc.) et expose les sols au ruissellement tout en conduisant à des conditions de topoinstabilité. Le départ de la forêt sur les versants occasionne l'arrivée massive des eaux dans la plaine.

¹ Source : 2ième RGPH en 1987

Ces eaux chargées de détrit, alimentent de façon brusque le réseau hydrographique qui subit des fluctuations suivant la fréquence des pluies en

engendrant les crues dont l'ampleur est fonction à la fois de l'intensité du ruissellement qui a eu lieu sur les pentes dénudées et du rythme des précipitations.

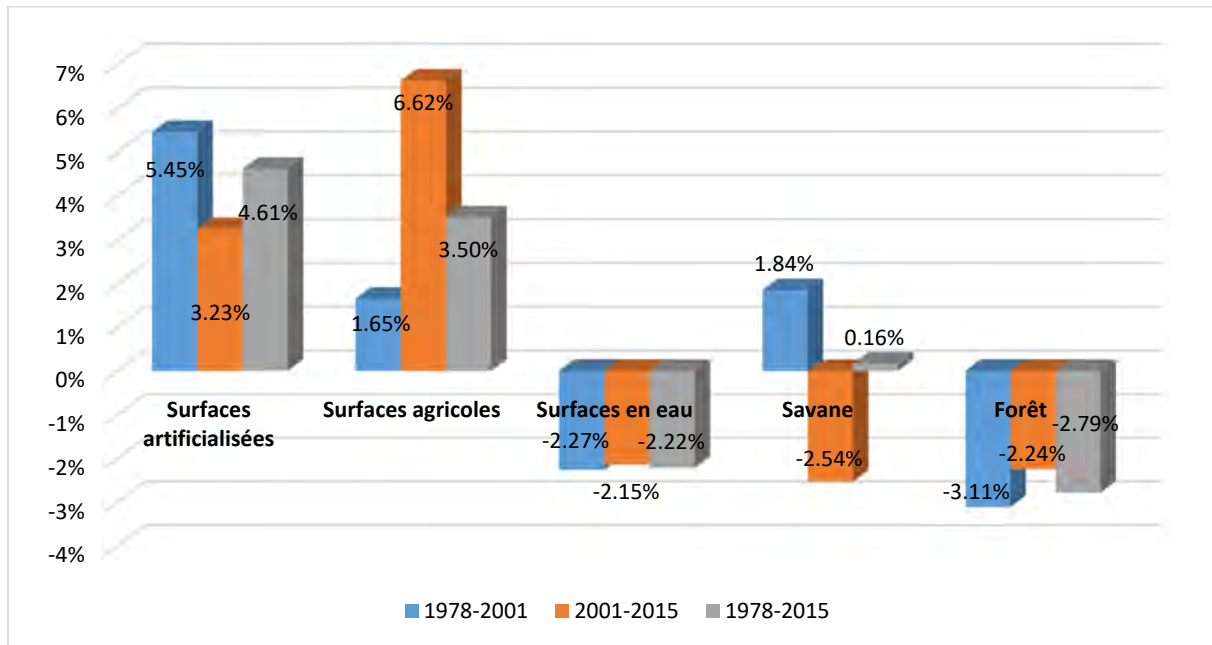


Figure 6 Variation des classes d'occupation du sol pour les différents pas de temps (1978-2001, 2001-2015 et 1978-2015)

3.2. Disparité spatiale de l'exposition des sols aux ruissellement

La susceptibilité des sols du bassin versant du Nkam au ruissellement a été réalisée grâce à la combinaison des facteurs ci-dessus présentés (Fig. 7). Le fait à constater sur cette carte est la variation des valeurs de CN trouvées.

Il apparaît une très grande variabilité des valeurs de CN dans le bassin étudié, traduisant ainsi la susceptibilité des sols au ruissellement. Les principaux facteurs contrôlant cette variabilité changent d'un site à l'autre : sur les sols de milieu naturel de l'escarpement de Foréké, c'est le degré de protection par le couvert végétal qui réduit le ruissellement (le CN varie de 25 à 64) ; sur les sols ferralitiques sur basalte à pente moyenne (plateau bamiléké), l'intensité du ruissellement dépend des pratiques culturales (culture avec labour et culture sans labour) ; sur les sols à fortes pentes mis en culture constamment (versant des monts Bambouto et Manengouba), c'est la situation du sol au cours du cycle cultural (déforestation et pratique régulière du labour qui ameublit les niveaux superficiels des sols)

qui est responsable des valeurs élevées de CN (> 82) ; sur les sols hydromorphes moyennement organiques de la plaine des Mbo, c'est la proximité de la nappe phréatique et la teneur en eau qui maintiennent les valeurs de CN élevées (> 82). En somme, l'hydrodynamisme des sols du bassin versant du Nkam dépend essentiellement de leur état de surface.

3.3. Un bassin versant susceptible au ruissellement dans son ensemble

L'application de la méthodologie décrite plus haut dans ses étapes 5 et 6 nous a permis d'estimer le coefficient de ruissellement du bassin versant. La superficie totale du bassin est de 2348,38 km², dont 204,86 km² est boisée sur sol A et possède un CN de 25. La pondération s'effectue en fonction de la superficie suivant la relation décrite à la dernière étape de la méthode. Le tableau 2 ci-dessous présente les résultats des combinaisons sol et utilisation du sol et le CN total du bassin. Il ressort de ce Tableau 2 que le CN moyen maximum est de 67 pour le bassin versant du Nkam. Avec cette valeur, l'infiltration commence à être faible, les ruissellements forts.

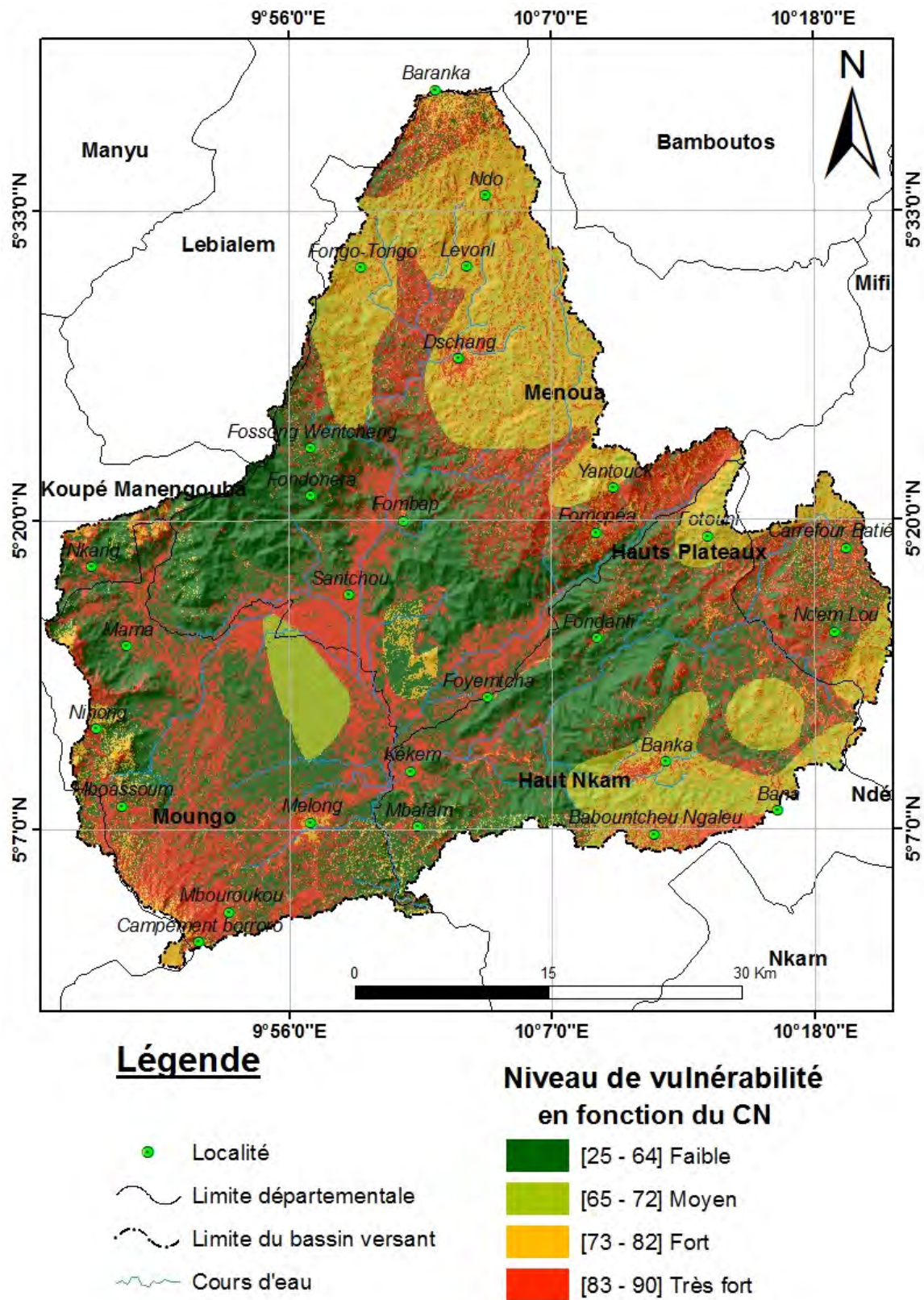


Figure 7 Image modélisée de la disparité spatiale de l'exposition des sols au ruissellement dans le bassin versant du Nkam, selon les combinaisons type de sol, utilisation du sol, pente et conditions hydrauliques

Table 2 Compilation du CN pour l'ensemble du bassin versant du Nkam

Combinaison surface boisée	Superficie km ²	Numéro de la courbe	Pondération
A	204,86	25	2,18
B	737,60	55	17,28
C	230,60	70	6,87
D	29,27	77	0,96
Total	1202,34		27,29
Combinaison surface agricole			
A	231,92	73	7,21
B	484,54	83	17,13
C	218,78	88	8,20
D	36,02	90	1,38
Total	971,44		33,92
Combinaison surface artificialisée			
A	60,92	65	1,69
B	97,66	76	3,16
C	15,85	84	0,57
D	0,17	88	0,01
Total	174,6		5,42
CN du Bassin versant du Nkam	2348.38		66,62 $\approx$ 67

Cette valeur du coefficient de ruissellement vient témoigner du recul des surfaces forestières au profil des surfaces agricoles et artificialisées. On peut constater à l'aide de la Figure 3 que les rapports conflictuels Homme-Environnement sont particulièrement marquant dans les milieux qui sont caractéristiques d'une fragilité naturelle, notamment la plaine des Mbo. Ces rapports sont à l'origine des inondations boueuses et l'envasement résultant de la formation de ruissellement érosif sur les parcelles agricoles situées en amont du bassin versant comme l'atteste la planche photographique 1.

Ces photos prises pendant les inondations du 04 août 2017 dans la plaine des Mbo (Echok et Santchou-ville) mettent en lumière les matériaux fluides et solides transportés par les eaux de ruissellement de l'amont vers l'aval.

4. Discussion

La méthode SCS Curve Number renseigne sur la capacité d'un bassin versant à ruisseler (Soulis et Valiantzas, 2012). Ce coefficient est fortement influencé par l'utilisation, le type et la pente des sols considérés. Les procédures d'analyse spatiale basées sur cette méthode pour l'appréhension de la formation du ruissellement de surface ont déjà fait

leur preuve à travers diverses études (Smith et Eggert, 1978; Hoang & Dumont, 1979; Yoo et al., 1993; Steenhuis et al., 1995; Bellall et al., 1996; Garen & Moore, 2005; Gassman et al., 2007; Trambly, 2012; Anh Tu Ngo, 2014). Ainsi, l'application de cette méthode dans le bassin versant du Nkam nous a permis de distinguer trois zones donc deux à susceptibilité élevée et une à susceptibilité faible. Ceci peut s'expliquer de diverses façons :

Le bassin juxtapose les basses (plaine des Mbo), moyennes (plateaux bamiléké) et hautes terres (monts Bambouto et Manengouba) à forte activité agricole. La pente du terrain est beaucoup plus prononcée sur ces deux dernières unités topographiques. Les pratiques agricoles observées sur ces unités topographiques homogénéisent le paysage et peut aboutir à l'apparition de vastes surfaces ruisselantes. La disparition du couvert végétale notamment sur les versants des monts Bambouto et Manengouba et autour des zones urbanisées, réduit les possibilités d'infiltration des eaux de pluie et facilite ainsi le transfert des eaux vers l'aval (ruissellement hortonien). En effet, en l'absence de toutes végétation, les filets d'eau sillonnent les parcelles labourées et enlèvent dans leur progression les éléments fins de surface, provoquant ainsi un

déblayage des sols. Les produits ainsi arrachés s'entassent en aval du bassin versants (plaine des Mbo) où la monotonie du relief limite l'écoulement des eaux entraînant par conséquent les inondations boueuses comme celles du 26 août 1978 (Tchoua 1984,1989) et tout récemment celle du 4 août 2017 à Echok par Santchou (Figure 8). Nos résultats

concordent avec ceux de Fonteh et al., (1998) qui, travaillant sur la culture des petits pois (*Pisium sativum* L.) dans les H.T.O.C, ont trouvé que pour une variation de pente de 0,89 à 2,69 % l'érosion augmenterait de 34 fois par rapport à la première pente.



Figure 8 Conséquences du ruissellement des eaux de pluie en aval du bassin versant (Plaine des Mbo)

En outre, la plaine des Mbo, de par son hydromorphie et la proximité de la nappe phréatique, les précipitations tombent sur une surface déjà saturée et ne peuvent que ruisseler. Ce ruissellement est fortement influencé par le cumul des précipitations qui provoque un gonflement progressif des nappes, ce qui sature d'autant plus rapidement le profil par le bas. On parle à cet effet de ruissellement sur sol saturé qui selon Ambroise (2004) est un phénomène efficace dans la génération des crues. Les secteurs ainsi saturés accompagnent fréquemment les cours d'eau, et le ruissellement ainsi généré contribue directement à l'écoulement de crue. C'est l'ensemble des raisons évoquées qui

laisserait indiquer des valeurs de CN beaucoup plus élevées dans certaines parties du bassin.

Cependant, le pourcentage boisé est généralement plus dominant sur certains sols en pente. Ces sols en pente sont communément moins sollicités pour l'agriculture en raison de la mécanisation difficile. Il est donc normal de constater une utilisation moins intensive du sol le long de l'escarpement de Foréké. L'effet d'une pente plus accentuée est donc compensé en bonne partie par une utilisation moindre du sol, ce qui explique le peu de variation des valeurs de CN aux alentours de la plaine Mbo. Les travaux d'Armand (2009), sur l'étude des états de surface du sol et de leur dynamique,

montrent que le couvert végétal retient, selon sa densité, sa nature et l'importance de la précipitation, une proportion variable de l'eau atmosphérique.

Toutefois, si certaines études menées sur le phénomène d'érosion dans les H.T.O.C. sont essentiellement descriptives (Tchawa, 1991, 1993; Fotsing, 1993 ; Tsayem Demaze, 1995), cet article va au-delà de cette description en évaluant le degré de sensibilité des sols du bassin versant au ruissellement. L'originalité de ce travail repose sur le fait, qu'outre la description, elle définit un espace différencier du risque d'érosion par ruissellement pour le bassin versant du Nkam à partir de la méthode SCS CN.

5. Conclusion

En somme, cette étude portait sur les sols dans le bassin versant du Nkam. L'objectif était d'évaluer à l'aide de la méthode SCS couplée au SIG la susceptibilité des sols au ruissellement dans cet espace intramontagnard. Au terme de ce travail, nous pouvons conclure sur la base des résultats obtenus que le bassin versant du Nkam a atteint un seuil inquiétant d'exposition des sols au ruissellement (CN=67). Cette situation est due au poids qu'exerce la régression du couvert végétal sur des sols pentus à fort usage agricole. En effet, il a été constaté une régression importante du couvert végétal en amont du bassin versant au profil des activités agropastorales, où le labour est couramment pratiqué. Environ 58% de la zone étudiée concerne les pentes fortes (> 15%) avec des sols constitués de minéraux d'un degré d'altérabilité assez élevé. On assiste à cet effet en saison des pluies au phénomène d'érosion par ruissellement responsable du lessivage des sols en amont qui conduit à l'engorgement et aux inondations boueuses en aval. Tous ces facteurs mis ensemble résument l'état actuel des sols du bassin versant du Nkam et par conséquent leur susceptibilité au ruissellement. À l'issue de cette étude, Il faudrait éviter d'attendre jusqu'au niveau où la remédiation nécessiterait des mesures curatives. Il est donc nécessaire voire primordial d'encourager les mesures correctives et préventives. Cet article approfondit ainsi la contribution de la géographie des risques sur la prise des décisions. L'étude de la susceptibilité des sols au ruissellement en milieu

tropical de montagne est un prisme à travers lequel, les problèmes liés au contexte de risques environnementaux sont mis en relief dans le but d'indiquer des pistes de solution pouvant réduire l'exposition des sols au ruissellement surtout dans les espaces tropicaux intramontagnards. La réduction de l'incidence du ruissellement peut donc se faire par l'introduction du reboisement des terrains caractérisés d'une nudité exceptionnelle et la pratique des techniques de culture sans labour et la construction des billons dans le sens des courbes de niveau sur les versants de colline. C'est dans cette perspective que nous invitons les instances nationales en charge des questions environnementales (MINEPDED), forestières (MINFOP) et agricoles (MINADER) et les populations locales pour une insistance dans le cadre des ODD (Objectifs du Développement Durable), notamment sur la préservation et la restauration des écosystèmes terrestres. La carte de susceptibilité des sols au ruissellement ainsi réalisée, peut cependant être un instrument de base à la prospection de cet espace géographique pour une meilleure affectation d'utilisation du sol.

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Les unités morphostructurales et pédologiques des monts Mandara face à la rétention en eau de surface

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RÉSUMÉ

Les monts Mandara constituent une unité morphologique particulière à l'Ouest de la région de l'Extrême-Nord du Cameroun. Cette singularité par rapport aux plaines bordières est liée à ses spécificités altitudinales, géologiques, géomorphologiques et pédologiques. Semblable à un ensemble de bourrelets rocheux enrochés aux formations superficielles régosoliques et lithosoliques minces et inexistantes par endroit, ce secteur est très souvent appréhendé comme un cadre hostile en matière de conservation d'eau, puisque les populations qui y vivent font face à un problème criard de son indigence en saison sèche. Ce problème d'eau ayant pour fondement les caractéristiques climatiques marquées par des sécheresses, les paramètres morpho-pédologiques caractérisés par la dissection du relief et la porosité des sols y sont très souvent associés. Son altitude surplombe toutes les plaines voisines avec une dénivellation de près de 500 m en moyenne. Cette partie caractérisée par des versants abrupts, des vallées compartimentant l'ensemble du paysage, la topographie à allure curviligne, le substrat rocheux à fleur de surface et des sols très poreux s'avèrent être des limites à la rétention naturelle (sans assistance) en eau de surface. Paradoxalement, grâce à cette forte dissection du relief avec des vallées importantes pour le socle du massif sain, peu recouvert par des altérites perméables et la présence des matériaux de base nécessaires à la construction des barrages, ce secteur pourrait offrir des possibilités d'aménagement de rétention artificielle d'eau de surface nécessaire dans ce milieu marqué par une saison sèche persistante (7 mois). Afin d'atteindre l'objectif qui est de montrer la capacité morphostructurale et pédologique des monts Mandara à retenir de l'eau en surface dans ce milieu semi-aride montagnard à substrat rocheux, la recherche des données secondaires, primaires et des analyses de laboratoire ont été effectuées. À l'issue de ces travaux, il ressort que les facteurs jadis contraignants sans intervention humaine revêtent un aspect positif pouvant se transformer en avantage pour la construction des ouvrages de rétention en eau de surface déjà très présents dans cette partie.

MOTS CLÉS barrage, eau, unité morphostructurale, rétention, pédologique, monts Mandara

ABSTRACT

The Mandara Mountains constitute a particular morphological unit in the west of the Far North region of Cameroon. This uniqueness compared to the border plains is linked to its altitudinal, geological, geomorphological and pedological specificities. Similar to a set of rocky ledges with thin regosolic and lithosolic surface formations that do not exist in places, this sector is very often apprehended as a hostile framework in terms of water conservation, since the populations who live there face a glaring problem of poverty in the dry season. As this water problem is based on the climatic characteristics marked by droughts, the morpho-pedological parameters characterized by the dissection of the relief and the porosity of the soils are very often associated with it. Its altitude overlooks all the neighboring plains with an average drop of nearly 500 m. This part characterized by steep slopes, valleys compartmentalizing the entire landscape, the curvilinear topography, the rocky substratum flush with the surface and very porous soils prove to be limits to natural retention (without assistance) in surface water. Paradoxically, thanks to this strong dissection of the relief with important valleys for the base of the healthy massif, little covered by permeable alterites and the presence of basic materials necessary for the construction of dams, this sector could offer possibilities for the development of artificial retention of surface water necessary in this environment marked by a persistent dry season (7 months). In order to achieve the objective which is to show the morphostructural and pedological capacity of the Mandara Mountains to retain surface water in this semi-arid mountainous environment with rocky substratum, the search for secondary and primary data and laboratory analyzes have been carried out. At the end of this work, it emerges that the previously restrictive factors without human intervention have a positive aspect that can turn into an advantage for the construction of surface water retention structures already very present in this part.

KEYWORDS dam, water, morphostructural unit, retention, pedological, Mandara Mountains

1. Introduction

Les monts Mandara, en raison de leur aspect massif, sont un ensemble découpé en trois unités morphostructurales : les bourrelets bordiers, les plateaux intramontagneux et les piémonts (Fig. 1). Il s'agit de la zone la plus assistée du Cameroun en matière d'aide et de projets de développement, c'est ce qui justifie d'ailleurs la mise sur pied par le Gouvernement de la MIDIMA¹. En fait, l'une des particularités de cet ensemble de hautes terres tient à sa population qui est parmi les plus pauvres, mais aussi l'une des plus denses du Cameroun (Gubry, 1988 ; 3e RGPH, 2010; MIDIMA, 2010).

Cette assistance est liée à de nombreux maux tels que la pauvreté, l'érosion des sols, l'amenuisement des ressources naturelles, l'insuffisance des équipements sociaux de base, le problème de divagation des bêtes et la question de l'accès à l'eau minent les populations de cette zone. De tous ces problèmes sus évoqués, celui de l'accès à l'eau se pose avec acuité ; pourtant ce secteur est le plus

pluvieux de l'Extrême-Nord. Le constat qui est fait dans ce milieu est qu'il y a un réel problème d'indigence en eau en saison sèche. Celui-ci se manifeste par l'accès difficile (longues distances, files d'attente interminables et promiscuité autour des points d'eau), la gestion compliquée de la ressource disponible et mobilisable, la surexploitation et la dispute de l'eau. Cette situation a poussé un riverain à dire que « *l'eau est plus rare à Mokolo² que l'argent* ».

Face à ce souci lié à une ressource incontournable à la survie et au bien-être des populations, de nombreuses stratégies d'adaptation face à ces difficultés d'accès à l'eau se sont développées et se multiplient encore afin de promouvoir le bien-être de la population. Cependant, le problème persiste et s'aggrave avec la population sans cesse croissante. Le climat en général et les paramètres géomorphologiques et pédologiques sont très souvent indexés comme facteurs premiers de cette vulnérabilité en eau (Beau-

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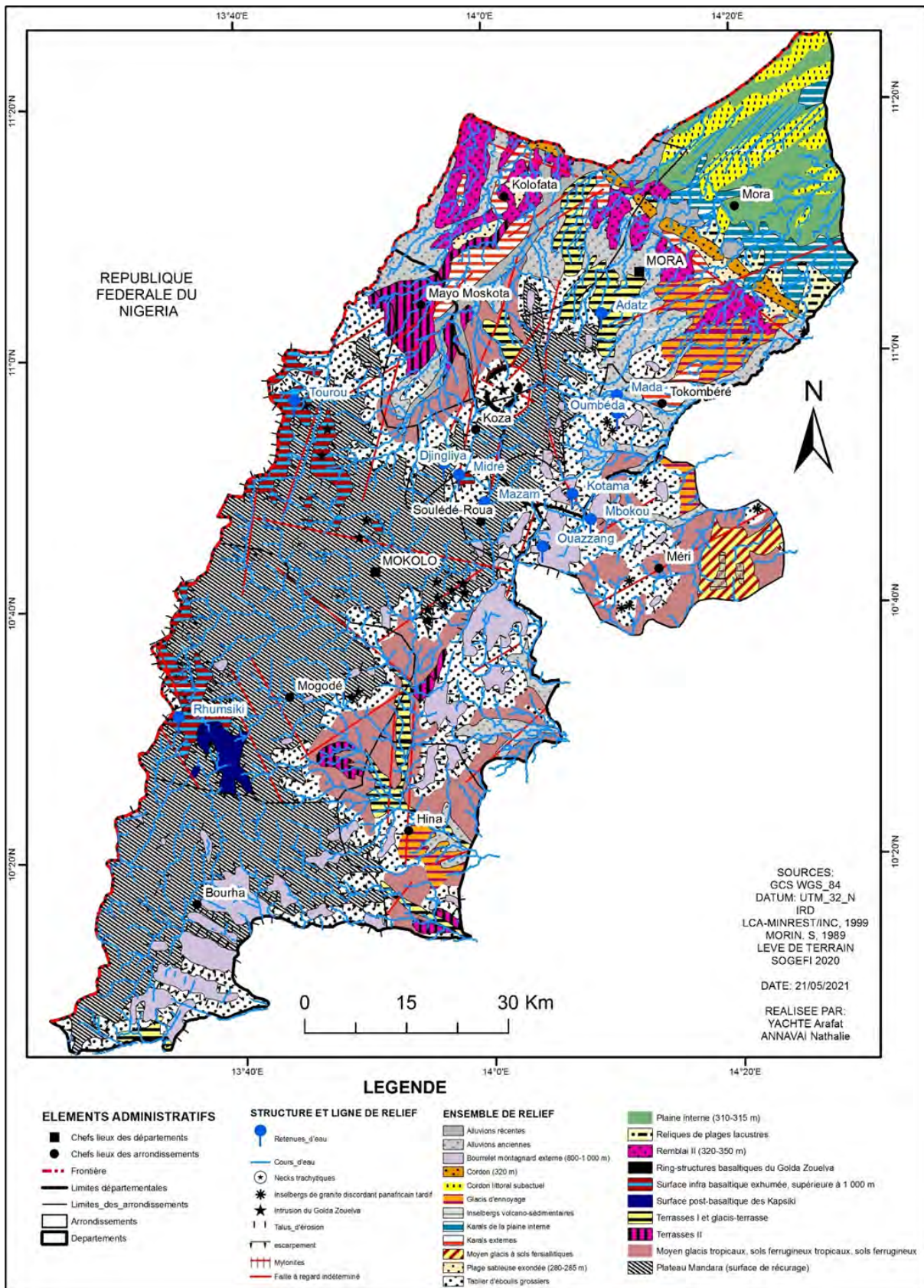


Figure 1 Géomorphologie des monts Mandara

vilain, 1989).

Par ailleurs, certains voient plutôt dans ce type de milieu d'autres avantages pour la valorisation des ouvrages de mobilisation des eaux de surface. Raison pour laquelle, le bureau d'étude français SOGREAH³ en 1976 a dressé une liste des sites de barrages à divers usages dans ce domaine, avec comme objectif de réduire la vulnérabilité naturelle. Cette idée est partagée par bon nombre d'auteurs (Hoogmoed et al., 1991; Critchley et al., 1991 et Wakponou, 2005) qui précisent que la conservation de l'eau dans de milieux pareils passe par la construction de petits barrages dans les vallées et la création des mares qui seront les lieux d'approvisionnement pour les hommes et d'abreuvement pour le bétail en saison sèche lorsque les puits et forages seront asséchés.

À travers ces 2 idées, on comprend qu'en dehors du climat, le relief est l'un des facteurs de vulnérabilité de rétention non assistée en eau de surface dans ce milieu. Toutefois, celui-ci peut se transformer en un avantage si l'on met des aménagements adaptés. Ainsi, il est question de répondre à la préoccupation de savoir : en quoi les caractéristiques géomorphologiques et pédologiques des monts Mandara présentent des avantages en matière de rétention en eau de surface ? Pour répondre à cette question, le présent travail a été structuré en 2 parties. D'une part, il s'agira d'analyser les éléments morphostructuraux favorables à la rétention en eau de surface. D'autre part, il s'agira de ressortir les prédispositions pédologiques qui concourent à la possibilité d'aménagement des ouvrages de rétention d'eau en surface et la description de ceux qui sont existants.

2. Matériels et Méthodes

2.1. Les données secondaires

Les ouvrages, les articles, les communications, les thèses, les mémoires, les rapports et les archives ont été utilisés dans le cadre de ce travail. Nous avons fait recours aux bibliothèques des universités de Ngaoundéré (FALSH), de Maroua (centre de documentation de l'ENS de l'université de Maroua, le

département de climatologie hydrologie et pédologie) et de la MIDIMA.

2.2 Les données primaires

2.2.1 Les observations directes et les levés de terrain

La collecte des informations directes par les observations de terrains a permis de découvrir le milieu naturel, de faire une mise au point sur le nombre de retenues d'eau construites dans les monts Mandara et de faire des levés de terrain pour mieux connaître la répartition spatiale des retenues d'eau.

2.2.2 Les prélèvements des échantillons de sol

Les prélèvements de 12 échantillons de sol pour des analyses de porosité, de la texture et de perméabilité se sont effectués sur les 20 premiers centimètres (horizon A).

2.2.3 Les enquêtes par questionnaire et les entretiens

L'enquête par questionnaire visait essentiellement les ménages. Un échantillon de 250 ménages auxquels le questionnaire a été administré dans 25 localités. Les entretiens avec les agents de la MIDIMA, les chefs de village, les gestionnaires de ces retenues d'eau et d'une vingtaine de focus groups avec la population, ont permis d'avoir des informations générales sur les dates de mise en eau, leurs buts, les volumes et les profondeurs.

2.2.4 Outils de collectes et de traitement des données

Des matériels de collecte (GPS, appareil photo, pelle-bêche, plastiques), les logiciels de traitement de texte (Microsoft Word), de réalisation des graphiques (Microsoft Excel), de réalisation des cartes (QGIS 3.12.3 et global mapper 19), ainsi que la réalisation des profils topographiques (Google Earth) ont été utilisés. Les échantillons ont été analysés à l'IITA de Yaoundé, à la FASA⁴ de Dschang et au Laboratoire de Géographie de l'ENS de Maroua.

3. Résultats

3.1. Les facteurs de prédisposition des sols au ruissellement

3.1.1. Des sols aux minéraux facilement altérables

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Du point de vue structural, les monts Mandara se caractérisent, par un relief plissé, métamorphisé et granitisé depuis les orogénèses précambriennes (Gazel, 1958). Ils s'inclinent du SW vers le NE et du NW vers le SE pour une direction structurale NNE-SSW à N-S ou subméridienne (Wakponou, 2005).

Une structure appartenant à la zone de socle héritée de la tectonique de soulèvement, les monts Mandara sont un ensemble complexe constitué de plateaux centraux, de sommets (collines, petites et moyennes montagnes) et de piémonts (Fig. 1). C'est la rupture des pentes brutales des bourrelets bordiers avec les plaines du pourtour qui lui confère cet aspect d'un ensemble montagneux (Fig. 2). Ces ensembles aussi distincts les uns des autres sont séparés par les escarpements et les talus d'érosion. Les secteurs pouvant être des atouts pour la valorisation des barrages sont les parties sommitales et les plateaux.

Les parties sommitales, encore appelés bourrelets bordiers (Wakponou, 2005), ils sont composés de roches granitiques et volcaniques (Wakponou, op cit) qui entourent les plateaux et dominant les plaines environnantes (Fig. 1). Ces sommets se composent de roches différentes aux altitudes différentes dont les plus hauts sont les hossérés Oupay et Ziver au Nord-Ouest de Mokolo; avec des altitudes respectives de 1492 et 1412 m. Ces parties sommitales aux allures différentes sont des zones au relief marqué avec des pentes abruptes (Fig. 2). Ce relief forme presque un cercle qui entoure les plateaux (Fig. 3).

Les plateaux intramontagneux, situés au cœur des montagnes, d'où leur appellation plateaux intramontagneux, sont constitués des roches métamorphiques (Wakponou, 2005) avec une altitude moyenne de 830 m (Fig. 4). Dans le détail, on y dénombre cinq plateaux. En fonction de l'altitude moyenne, du bas plateau au plus haut plateau, on distingue les plateaux de : Soulédé-Roua (750 m) ; de Bourrah (800 m) ; de Mokolo et de ses environs (800 m) ; Kapsiki (900 m) et de Mabourki (900 m).

Au sud de l'arrondissement de Bourha en passant par Mogodé-Mokolo-Roua en remontant par Kouron, se hisse un relief accidenté avec une altitude maximale de 1 030 m au neck de Rhumsiki et une minimale de 500 m au nord du plateau de Kouron

(Fig. 5). Soit une dénivellation d'environ 530 m. Ces chiffres attestent que nous sommes en présence d'un relief plutôt accidenté. Ces profils en travers des 5 plateaux des monts Mandara (Fig. 4), donnent un aperçu sur la déclivité des terrains constitués de creux et de sailli avec des vallées aux déclivités allant de 100 à 530 m au sein et entre les plateaux. Ces dépressions sont traversées par des mayos dans lesquels se collectent toutes les eaux de précipitation. Ces versants escarpés favoriseraient la rétention en eau de surface qui peut être retenue par une digue grâce à la forte dissection du terrain.

Globalement, le relief de la zone des monts Mandara forme un gigantesque ensemble qui se caractérise par ses aspects massifs en dents de scie. Cet ensemble marqué par des vallées qui entrecoupent ce bombement de socle lui confère un relief accidenté. Ces caractéristiques sont des atouts pour le blocage de l'eau dans les vallées.

La morphologie des monts Mandara se présente d'une manière générale comme une succession de montées et de descentes formant des continuités de vallées. Selon Carrère (1994), « un site de barrage, au sens topographique, se place sur un verrou, resserrement de la vallée situé juste en aval d'une cuvette naturelle susceptible, une fois fermée, de constituer un réservoir de volume suffisant ». Le relief de ce secteur est disséqué par les cours d'eau logés dans des vallées. Ces dernières, aussi nombreuses que diversifiées sont des atouts incontestables pour l'édification des barrages (Fig. 6).

Son relief comprenant de nombreuses vallées, constituent des obstacles naturels et facilitent l'implantation des digues. Ces vallées constituent naturellement des dépressions et des cuvettes pour la rétention des eaux qui devront être barrées par des digues. Ce type d'environnement à site étroit, précédé à l'amont par un élargissement de la vallée se localise partout dans les Monts Mandara (Fig. 7). Les vallées sont semblables aux entonnoirs dans lesquels les eaux se dirigent rapidement vers l'exutoire. En fait, ce type de vallées est idéal pour l'emplacement, économique et nécessite moins de temps et d'effort.

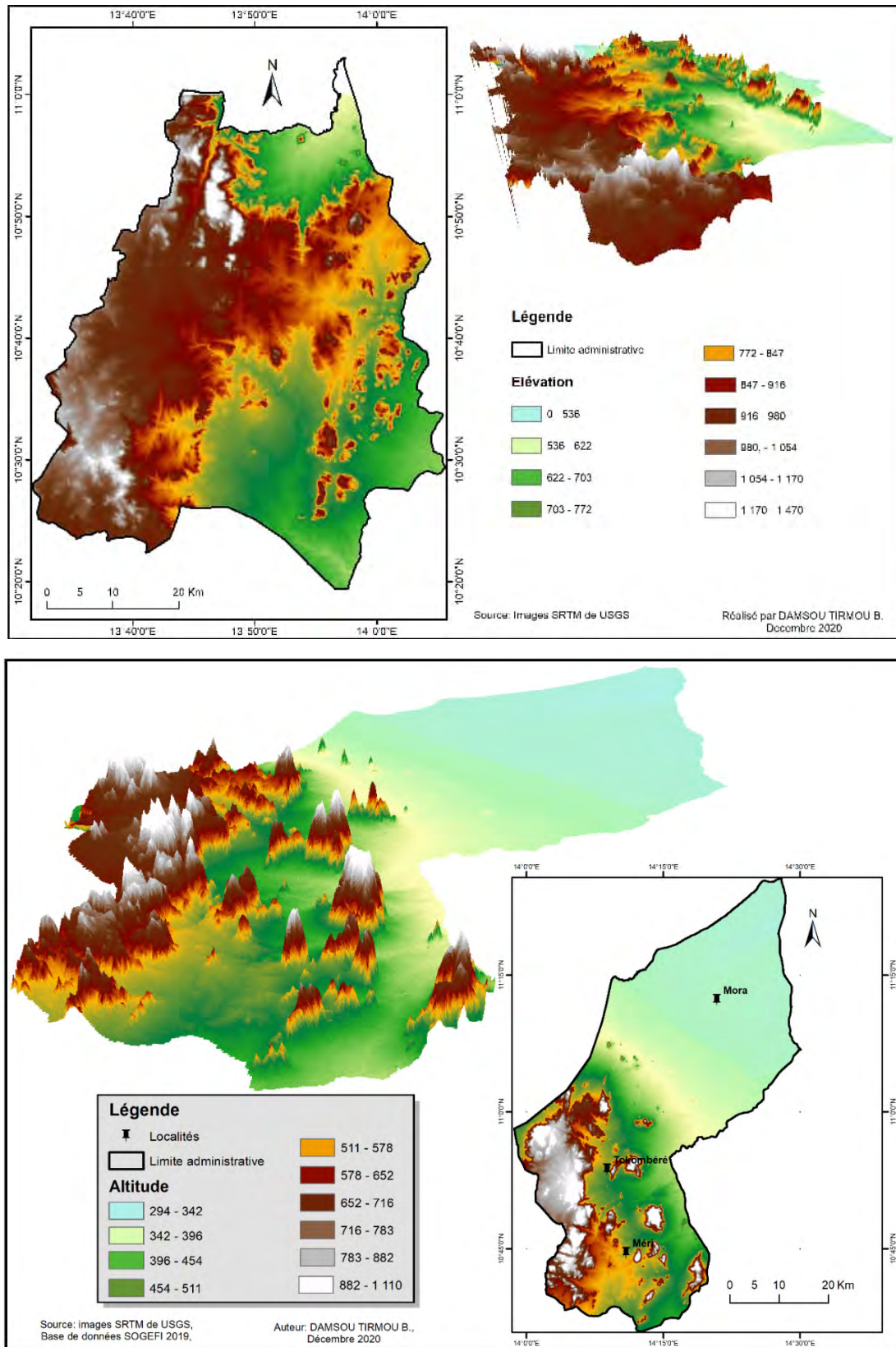


Figure 2 Modèle Numérique de Terrain (MNT) des secteurs des monts Mandara: haut le secteur du sud et bas le secteur du nord



Figure 3 Dôme rocheux à Tala Zoulgo

3.1.2 Le socle et des blocs rocheux offrant de nombreuses possibilités de points de construction des barrages

La constitution d'une retenue d'eau requiert du socle cristallin avec des propriétés minimales favorables en matière d'étanchéité naturelle. Dans les monts Mandara, l'étanchéité⁵ naturelle des roches du soubassement n'est pas à démontrer. En effet, le socle cristallin est formé essentiellement de gneiss associés aux granites dans les montagnes qui sont des roches imperméables. Les reliefs granitiques divers constituent le bourrelet montagneux bordier disposé tout autour des plateaux et les paysages métamorphiques sont ceux des plateaux intérieurs qui sont pour l'essentiel recouverts de roches métamorphiques et de quelques roches granitiques et volcaniques (Wakponou, 2005) (Fig. 5).

Tous ces plateaux dont les altitudes moyennes varient entre 750 et 900 m sont constitués de roches massives fracturées localement. Il en est de même pour les secteurs montagneux où parfois aucun manteau d'altérites ne recouvre les roches. Ce socle est un avantage pour la construction des barrages dont l'endiguement sera moins coûteux. Car, lorsque le substratum est sain (rocher ou terrain compact) comme c'est le cas dans les monts Mandara, cela ne nécessite pas de fouilles excessives qui s'avèrent

coûteuses et demandant assez de temps. De plus, ces roches sont utiles pour la construction des barrages et ouvrages annexes.

De nombreux versants des montagnes et collines se distinguent par des affleurements de blocs de granite. C'est la raison pour laquelle Morin (2000) a qualifié certains secteurs des monts Mandara de « monde de pierraille ». De blocs rocheux à perte de vue se font remarquer dès l'amorçement des monts Mandara qu'importe le chemin emprunté. Une citadelle de blocs grenus structure ce paysage pittoresque (Fig. 8). Ces blocs rocheux constituent des matériaux de construction des digues et les ouvrages annexes.

En effet, la disponibilité des matériaux nécessaires et utilisables pour la construction d'un barrage sur le site a une incidence considérable, voire prépondérante sur le choix du site d'implantation des retenues d'eau et du type de retenue d'eau. Des roches cristallines massives recouvrent toute la zone et peuvent être utilisables pour l'enrochement, le remblai ou protection des talus des digues. Les agrégats à béton comme les grains de quartz (sable) provenant de l'altération des roches grenues plutoniques et métamorphiques peuvent servir de

⁵ Étanchéité: L'imperméabilité des roches aux liquides, aux gaz

liant (ciment) pour former un béton afin de construire les digues et les ouvrages annexes.

Tous ces matériaux disponibles sont de véritables aubaines, qui pourront être utilisés pour la construction des retenues d'eau de ce secteur. La possibilité d'extraire ces matériaux de l'emprise de la retenue, permettra également d'accroître la capacité de stockage. Cela minimiserait les coûts de transport et d'achat des matériaux dans les zones d'emprunt. Puisque les matériaux locaux in situ seront disponibles et valorisés.

3.2 Les prédispositions pédologiques à la construction des retenues d'eau

3.2.1 Des formations superficielles très poreuses, mais peu profondes facilitant la réalimentation des nappes souterraines

Les sols des monts Mandara (Fig. 9), en raison de leurs textures sablo-limoneuse et sableuse et de leur structure peu compacte sont des sols très poreux. Cependant, ces sols peu profonds ne posent pas problème pour la rétention de l'eau en surface.

La composition texturale des sols des monts Mandara telle que présentée au Tableau 1 met en évidence les différents constituants minéraux des sols qui sont constitués entre 74,74 et 86,82 % de sable. Vu la proportion de sables, limons et argiles, il

Tableau 1 Les caractéristiques granulométriques des sols

Échantillon	Sable (%)	Limons (%)	Argile (%)	Classe texturale
Rhumsiki	74.74	13.42	11.83	Sablo-limoneux
Mada	74.82	15.35	9.83	
Oumbeda	78.67	15.42	5.9	
Tourou	80.74	11.42	7.83	
Kotama	80.96	13.21	5.83	
Adatz	81.03	13.28	5.69	
Mazam	81.03	11.14	7.83	
Mbokou	82.89	9.28	7.83	Sableux
Djingliya	82.96	11.21	5.83	
Ouazzang	86.67	9.42	3.9	
Midré	86.82	9.28	3.9	Sablo-limoneux
Moyenne	81.03	12.03	6.92	

Source : Analyse de laboratoire IITA Yaoundé.

s'agit majoritairement des sols à texture sablo-limoneuse pour une moyenne de 81,03 % de sables, 6,92 % d'argiles et de 12,03 % de limons.

La porosité texturale des sols des monts Mandara a été mise en évidence grâce aux travaux de Bédard (2014) (Tableau 2). Ce qui a permis de ressortir les classes texturales auxquelles appartiennent les sols des monts Mandara. Ces sols sont de granulométrie moyenne (0,465 - 2 mm). De ce fait, les sols à granulométrie moyenne ont une porosité utile qui est importante (27,5 - 32,5 %) ce qui témoigne du taux de vide significatif entre les matériaux. Ils sont caractérisés par une porosité comprise entre 36 et 59 % pour une moyenne de 49,09 % des proportions de vide entre les particules. Ceci est réel puisque la porosité de rétention en eau est faible (15-10 %). La structure de ses sols lui confère une bonne capacité d'infiltration.

La perméabilité tout comme la porosité sont fonction de la texture et de la structure des sols. On constate que la vitesse de perméabilité diminue avec la taille des matériaux, il y a donc une forte corrélation positive entre la taille des grains et la vitesse de perméabilité. Les sols des monts Mandara appartenant aux 2 classes texturales (sableux et sablo-limoneux), ont de ce fait les valeurs les plus élevées comprises entre 210 et 61 mm/h. Ce qui dénote d'une bonne perméabilité.

Pour mieux mettre en exergue cette perméabilité, 100 mg de tous les sols des monts Mandara et 50 ml d'eau ont été analysés afin de vérifier. La masse volumique (densité apparente) moyenne d'un solide est de 2,65 g/cm³. Les valeurs plus élevées de la densité apparente signifient que les vides sont réduits et que les particules sont fortement compactées et participent au ralentissement du processus d'infiltration. Par contre, les valeurs plus basses signifient une abondance des vides, les particules moins compactées et il y a une bonne infiltration (Tableau 3). Ceux des monts Mandara ont des valeurs basses ce qui signifie qu'ils retiennent moins d'eau. À l'issue des analyses, il apparaît que l'eau des échantillons de sols des sites des monts Mandara qui sont sablo-limoneux et sableux se filtrent rapidement par rapport à celui de la terre de Karal qui est argileuse. Car il y a un écart de plus de 2 h de temps entre les écoulements d'eau des sols sablo-limoneux et sableux et ceux issus des sols argileux. D'où la plus ou moins grande perméabilité des sols des monts Mandara.

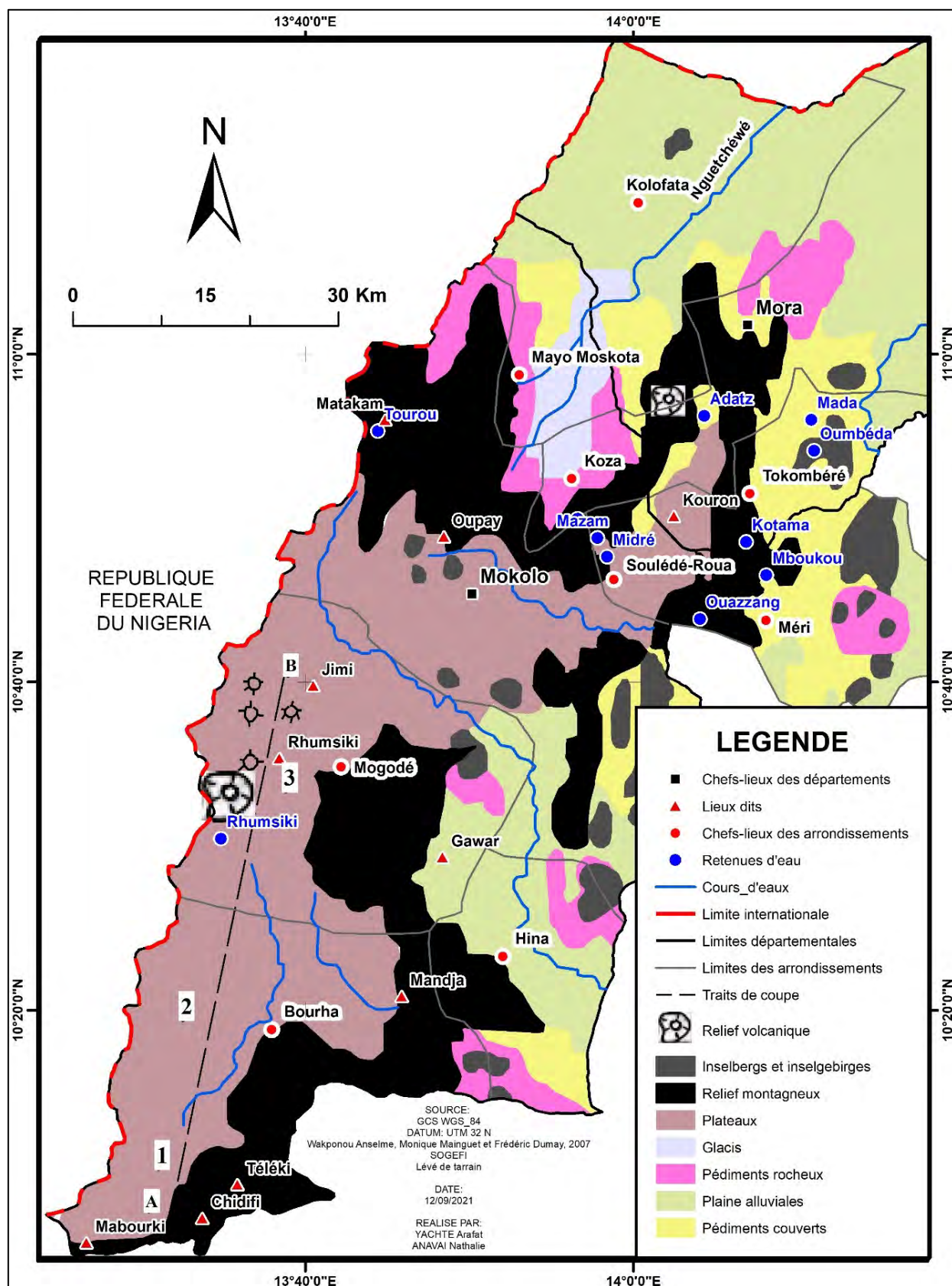
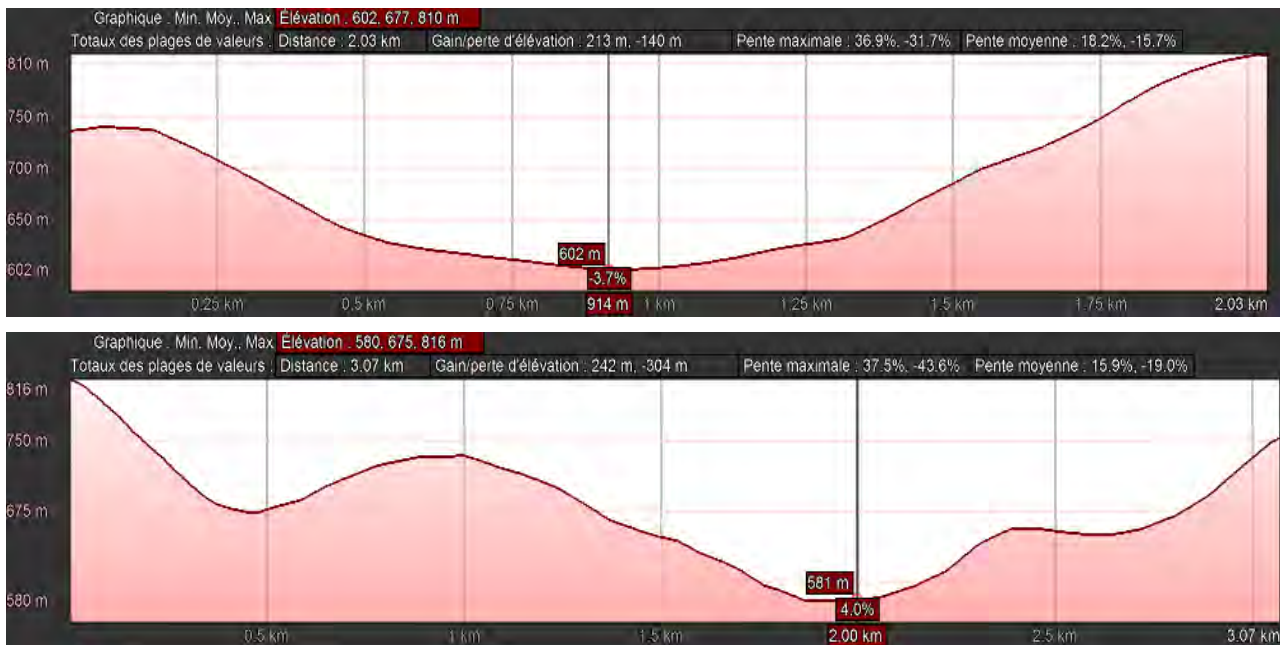
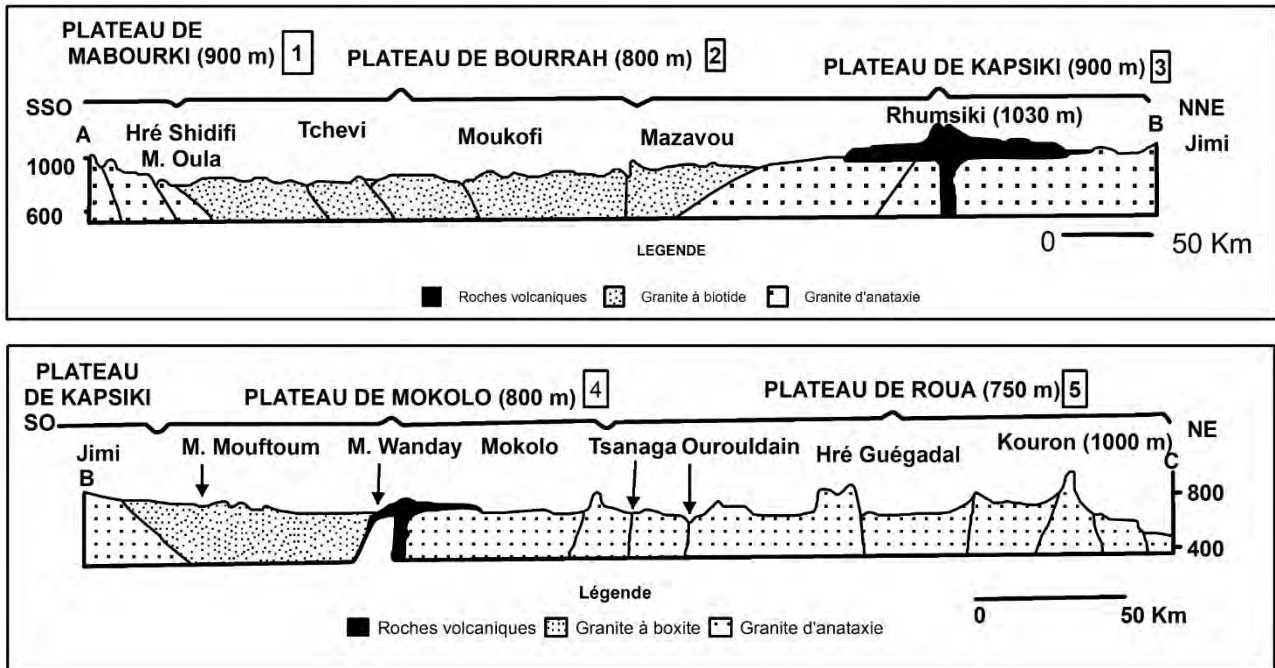


Figure 4 Carte géologique des monts Mandara



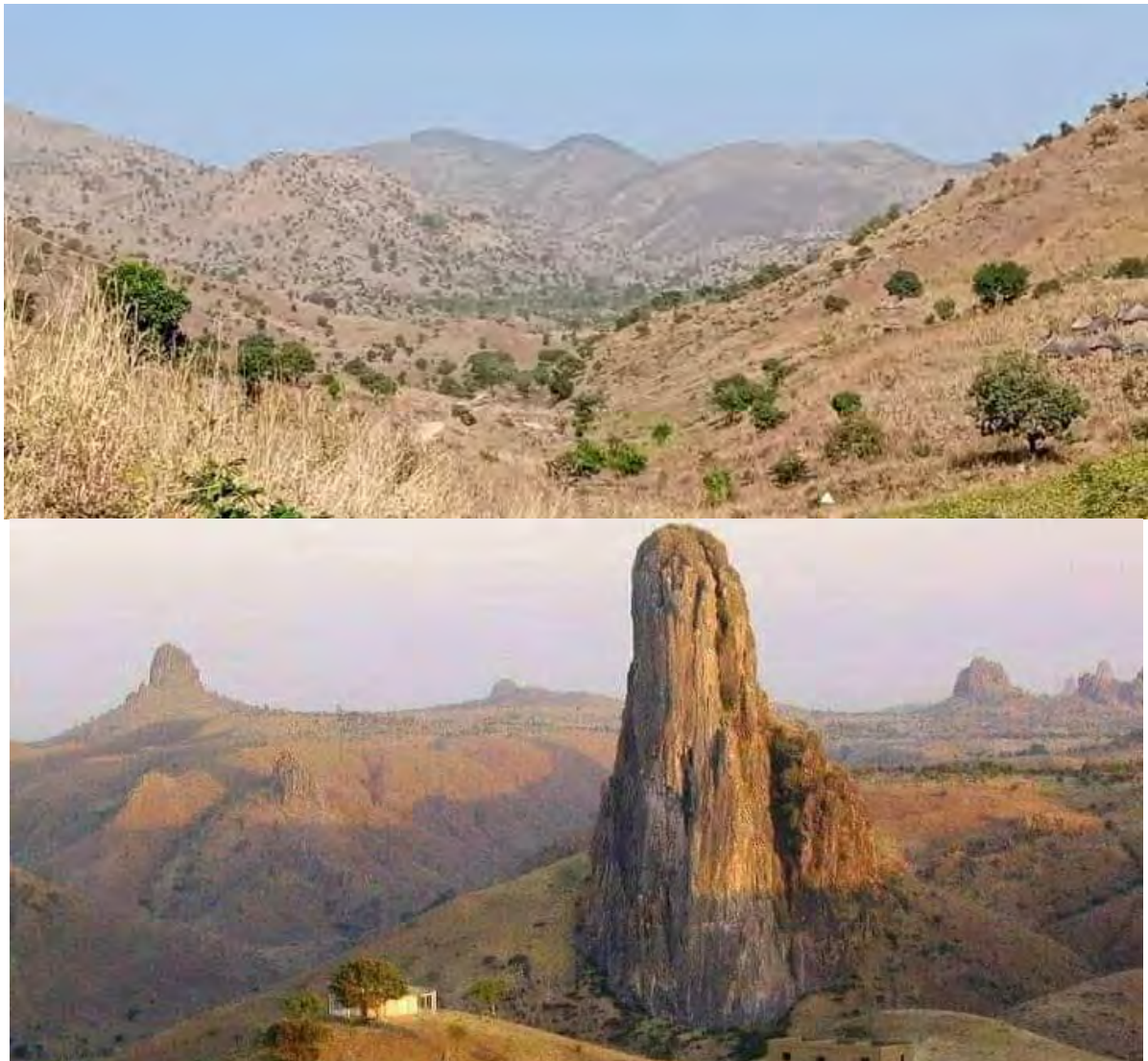


Figure 7 Les traits morphologiques des monts Mandara

Les sols des monts Mandara étant perméables (Tableaux 3 et 4), sont un avantage pour la recharge des nappes qui réalimenteront les puits et forages majoritairement asséchés en saison sèche. En effet, selon les enquêtes de terrain, la population utilise à 80 % les ouvrages de mobilisation des eaux souterraines pour la satisfaction de leur besoin. Cette capacité de recharge des nappes via des sols poreux serait un avantage incontestable pour ces secteurs, puisque ses barrages très souvent construits ne sont toujours pas accompagnés de stations de potabilisation (Tableau 5).

Les sols des monts Mandara, bien qu'étant poreux et perméables ne sont pas des freins à la construction des barrages et à la rétention de l'eau

en surface puisqu'ils sont peu profonds. Pour Westphal et al. (1981), ces sols sont sur alluvions présentant des caractéristiques suivantes : bien drainés, peu profonds, forte pente, caillouteux, recouverts de minéraux bruts squelettiques ou uniquement de rochers nus et d'arènes (Fig. 4). Ils sont caractérisés par un horizon A avec une profondeur de 0-20 cm, un horizon B très souvent absent et un horizon C de 20-50 cm (Brabant et Gravaud, 1985) qui repose sur la roche mère. En effet, ces sols sont minces (50 cm d'épaisseur), et cette épaisseur réduite ne pose pas véritablement problème pour l'emmagasinement de l'eau en surface via les barrages puisqu'ils se gorgent rapidement d'eau ce qui permettrait leur maintien en

Tableau 2 La porosité des sols des monts Mandara par rapport à leur classe granulométrique

Classe granulométrique	Dimension (mm)	Sites	Porosité utile (%)	Porosité de rétention (%)
Argile	<0.02		< 10	> 40
Limon	<0.062		10-15	30-35
Sable fin	0.062 – 0.465	Adatz Mazam Rhumsiki Ouazzang Kotama	15-30	315-30
Sable moyen	0.465 - 2	Midré Tourou Djingliya Mbokou	27.5-32.5	10-15
Sable grossier	2 – 4.75	Mada Oumbéda	30-32	7.5-10

Source : Bédard (2014) adapté par Annavaï Nathalie.

Tableau 3 La densité apparente des sols et capacité d'infiltration

Grand secteur	Site de prélèvement	Da (g/cm ³)	Temps de filtration et d'écoulement (minutes)	Quantité d'eau filtrée sur 50 ml
Monts Mandara	Adatz	1.09	45	30
	Rhumsiki	1.12		
	Midré	1.12		
	Oumbéda	1.21		
	Mazam	1.24		
	Mbokou	1.31		
	Tourou	1.41		
	Djingliya	1.56		
	Kotama	1.59		
	Mada	1.61		
Maroua	Ouazang	1.7	180	17
	Kongola	2.6		

Source : FASA, Unité de Recherche d'Analyse de Sols et de Chimie de l'Environnement (Université de Dschang) et Annavaï (2013).

Tableau 4 Taux d'infiltration détaillé en rapport avec le type de sol

Type de sol	Conductivité hydraulique saturée (mm/h)	Possibilité d'infiltration	Caractéristiques des sols des monts Mandara
Sableux	210	Bonne	Sableux
Sable limoneux	61	Bonne	Sablo-limoneux
Limon sablonneux	26	Moyenne	-
Limon	13	Moyenne	-
Limon silteux	6.8	Faible	-
Limon argilo-sablonneux	4.3	Faible	-
Limon argileux	2.3	Faible	-
Limon argilo-silteux	1.5	Faible	-
Argile sablonneuse	1.2	Faible	-
Argile silteuse	0.9	Nulle	-
Argile	0.6	Nulle	-

Source : Adapté de Ferguson, 1994 et Source. Musy et Soutter, 1991.

Tableau 5 Les caractéristiques des retenues d'eau réalisées dans les monts Mandara

Dénomination des retenues	Années de réalisation	Volume	Profondeur à la digue	Surface à plein de la retenue d'eau (ha)	Buts principaux	Type de barrage	Surface à plein de la retenue d'eau (ha)	Type et classe des barrages
Mokolo-Douvar	1978	5 000 000	25	85	Adduction en eau potable	Poids	85	Grand (A)
Chidifi	1984	5 000 000	25	100	Adduction en eau potable	Poids	100	Grand (A)
Tourou	1984/1990	900 000	12	6	Adduction en eau potable	Poids	6	Moyen (B)
Oumbéda	1980/1982	144 000	12	3.46	Adduction en eau potable	Poids	3.46	Petit (C)
Adatz	1983/1985	85 000	7	0.83	Stockage	Poids	0.83	Petit (C)
Ouazzang	1988/1990	75 000	9.5	1.09	Adduction en eau potable	Poids	1.09	Petit (C)
Mogodé	2012	75 000	10	2	Adduction en eau potable	Poids	2	Petit (C)
Kotama	1978/1980	47 000	8	1.76	Adduction en eau potable	Poids	1.76	Micro (D)
Hodogo	1988	45 000	8	1	Stockage	Poids	1	Micro (D)
Djingliya	1972	30 000	9.5	1.40	Stockage	Poids	1.40	Micro (D)
Mbokou	1978/1980	22 000	8.15	0.92	Adduction en eau potable	Poids	0.92	Micro (D)
Mazam	1987/1989	20 000	7.70	0.18	Adduction en eau potable	Poids	0.18	Micro (D)
Midré	1987/1989	15 000	6.50	0.17	Stockage	Poids	0.17	Micro (D)
Rhumsiki	1988/1989	10 000	9	0.6	Stockage	Contre-fort	0.6	Micro (D)
Mada	1988/1989	10 000	7.17	0.15	Stockage	Poids	0.15	Micro (D)

Source : Mission des barrages de la région des Monts Mandara (mai 1998) et travaux de terrain en 2016.



Figure 8 Affleurement des roches granitiques sur les versants des montagnes

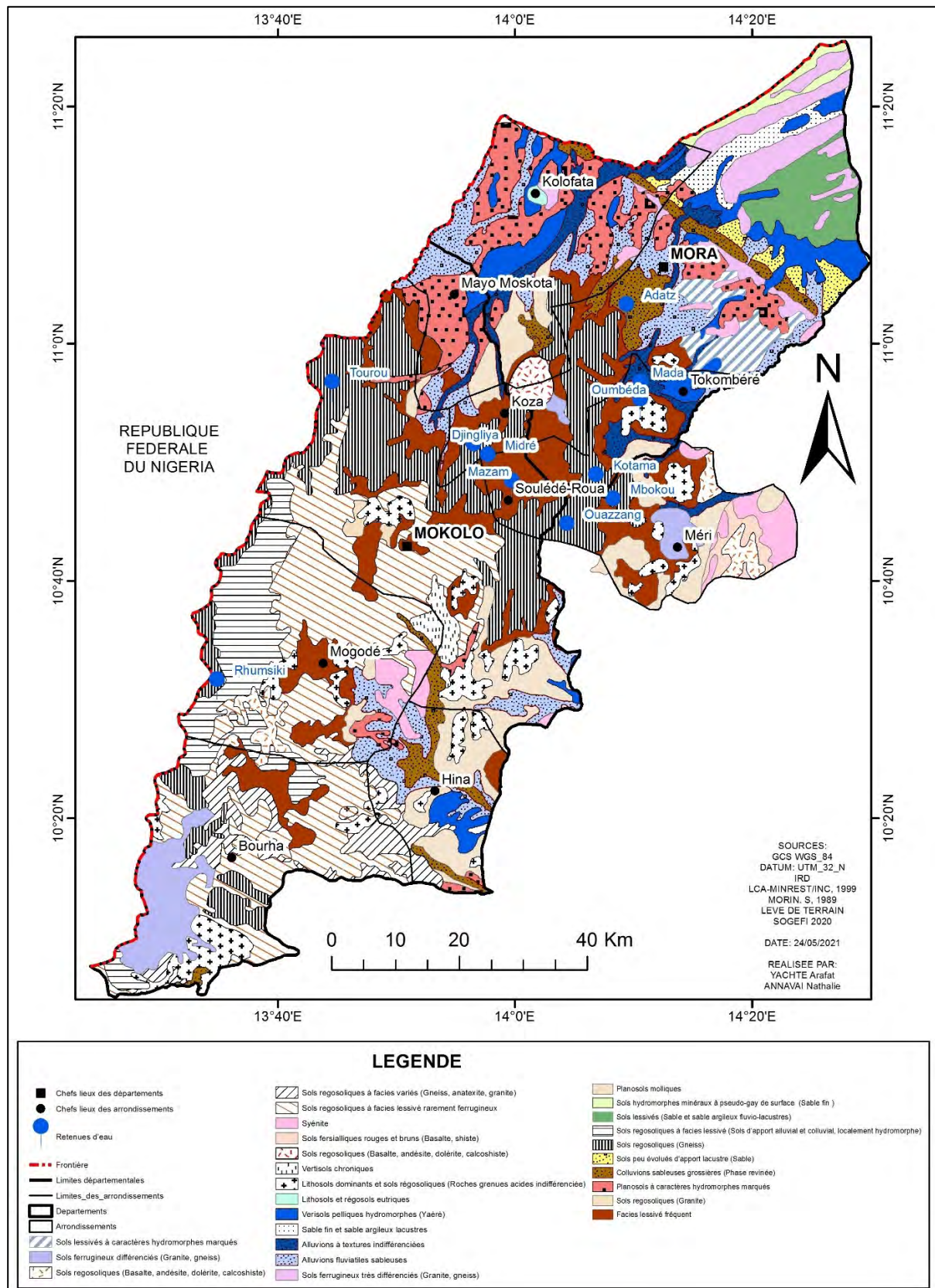


Figure 9 Carte pédologique des monts Mandara

surface après le blocage par des digues. De plus, ces sols sont très souvent absents et des dalles affleurant par endroit constituent déjà un soubassement stable pour les cuvettes des retenues d'eau.

Les possibilités à l'édification des barrages dans les monts Mandara sont déjà des réalités sur le terrain. En effet, ce secteur est la seule zone au Cameroun où l'on dénombre plus d'une dizaine de barrages.

3.2.2 Inventaire et caractéristiques des barrages des monts Mandara

Les avantages qu'offrent les monts Mandara en matière de construction des retenues d'eau avaient déjà été compris. En effet, les populations locales depuis les années 30 construisaient des retenues d'eau artificielles (Seignobos et Iyebi-Mandjeck,

2000). De nos jours, dans cet ensemble, on dénombre 15 barrages de retenues d'eau moderne avec des volumes compris entre 5 000 000 et 10 000 m³ d'eau (Tableau 5). Ils ont été construits entre 1972 à 2012 et se localisent entre 105 082 et 109 747 de latitude Nord et 135 837 et 141 318 de longitude Est dans 3 Départements et 9 Arrondissements (Fig. 1). Ils s'intègrent de façon naturelle dans l'environnement sans grand effort (Fig. 10). Constitués en majeure partie des barrages-poids dont les digues sont à béton cyclopéen⁶, ces ouvrages sont à usage multiples (Tableau 5). Les barrages permettent de redonner vie aux ouvrages comme les puits et les forages qui s'asséchaient complètement en saison sèche avant la construction de ceux-ci. Ils sont d'un avantage incontestable pour les populations et leur bétail.



Figure 10 Les barrages intégrés naturellement dans le paysage

⁶ Béton contenant de gros blocs de pierre, des moellons, des galets

4 Conclusion

Les monts Mandara, longtemps perçus comme une zone de frein pour la disponibilité en eau de surface du fait de ses paramètres morphostructuraux et de ses caractéristiques pédologiques. Ces mêmes caractéristiques analysées sous l'angle des aménagements anthropiques de rétention de ladite ressource, offrent des possibilités importantes à la rétention en eau de surface au niveau des vallées. Cet autre regard sur l'aspect de caractéristiques suscitées est lié à la transcendance de ces données stationnelles qui sans intervention humaine, demeure des facteurs contraignants à la disponibilité de la rétention en eau de surface. C'est alors que la solution de rétention artificielle de l'eau en surface, développée depuis la période coloniale par certaines populations s'est vue renaître et remise en surface depuis le début des années 70 jusqu'en 2012, et de nombreux projets de construction sont d'actualité pour le bien-être des populations. Cependant, ces ouvrages implantés dans cette zone réputée pour ses processus dynamiques accélérés, qu'en est-il de leur durabilité?

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Rock glaciers vegetation colonization in Retezat Mountains: implications for morphodynamics and palaeoclimate reconstruction

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ABSTRACT

Vegetation is an important ecological indicator of rock glaciers' activity as its development marks the transition from active to inactive and relict states. Most rock glaciers in Retezat Massif are covered by vegetation, mostly *Pinus mugo* shrubs but also herbaceous and forestry vegetation. In this paper, 82 rock glaciers from Retezat are classified into six classes according to vegetation characteristics, rock glaciers' vegetation changes are assessed using old aerial and ground imagery, and the timing of shrubs vegetation growth is investigated by means of dendrochronology methods applied on three rock glaciers. The study's main purpose is to identify what caused the vegetation colonization, climate variation, or rock glacier inactivation. Results indicate that altitude is the dominant factor of rock glacier vegetation cover variability. In most cases, the vegetation patches increased in density and only slightly in surfaces. The last rock glaciers' vegetation colonization initiated in the 18th century, beginning with 1830s on all the three rock glaciers probably after the last local Little Ice Age cold spell. We conclude that the investigated rock glaciers were pseudorelict during the entire Holocene and that recent vegetation settling is not the first one in the Holocene. However, the scenario of rock glaciers inactivation that caused vegetation growth, previously advanced in the literature, cannot be excluded, especially in light of the recent discoveries of rock glaciers activity, and further investigations should be done to come to a less ambiguous outcome.

KEYWORDS rock glaciers, *Pinus mugo*, climate change, subalpine vegetation changes, Retezat Mountains

1. Introduction

Rock glaciers are harsh environments for vegetation colonization because of their openwork scree

structure that drain water efficiently, hinder fine material accumulation and soil formation, and favor snow persistence (Canone and Gerdol, 2003; Tampucci et al., 2015). As active rock glaciers can

sporadically allow lichens and plants adapted to surface blocks displacements, the inactive and relict ones have a more extensive and diverse vegetation cover favored by stable ground and more availability of areas with fine debris and soil (Burga et al., 2004). This observation allowed using the vegetation cover as an empirical indicator for the activity status of rock glaciers (e.g., Scotti et al., 2013; Onaca et al., 2017; Popescu, 2018; Jones et al., 2019). Fine materials are usually found on rock glacier front and lateral edges, which are generally less mobile than the rest of the body and are therefore the first to be colonized by vegetation. Lichens develop extensively on rock glaciers, even on the active ones, and that is possible because of the passive movement of gelifractions that remain exposed to daylight with the same facets. Moreover, previous studies showed that *Rhizocarpon geographicum* agg. thalli are well correlated with the rock glacier age and that lichens cannot grow in areas with snow persistence of more than nine months in a year (Burga et al., 2004). Overall, the topoclimate, surface movement and fine sediments availability are the main controlling factors in vegetation distribution on rock glaciers.

Rock glaciers are landforms sensitive to climate changes and air temperature rising induces mostly dynamics acceleration, but as a final consequence, it can also inactivate them due the ice content reduction or disappearance (Fleischer et al., 2021). Reducing activity rates can increase floristic diversity on rock glaciers (Cannone and Piccinelli, 2021).

Several species of trees can be used as palaeoenvironmental indicators applying dendrochronology (Pearl et al., 2020). By analyzing tree rings one can determinate age, palaeoclimate changes or occurrence/variability in geomorphologic processes interacting with the growing plants. Since rock glaciers are mainly located above the treeline, the mostly used species in tree ring studies, i.e. spruce trees, are unavailable. Thus, alpine shrubs are the only ones that can offer palaeo-environmental information. Recent studies revealed that alpine shrubs like *Salix helvetica* (Gärtner-Roer et al., 2013) and *Linantus pungens* (Franklin, 2013) are sensitive to and can archive information about rock glacier activity or climate variation. Other species with a large occurrence above the tree line in the European

Mountains and Carpathians, in particular, is the dwarf pine (*Pinus mugo*), a common shrub belonging to the subalpine vegetation belt (Šibík et al., 2010), widespread also on relict rock glaciers (Kellerer-Pirklbauer et al., 2015; Popescu, 2018). Dendrochronology was applied on *Pinus mugo* in mountain areas (Palombo et al., 2010; Timiş and Popa, 2011; Dai et al., 2017) but rarely on rock glaciers so far (Urdea, 1998). Thus, the current paper aims at 1) presenting the ages of *P. mugo* shrubs located on rock glaciers and in their vicinity; 2) discussing the relevance of *P. mugo* ages in relation to rock glaciers dynamics and 3) assessing the rock glacier vegetation cover in Retezat Mountains and changes in the last 100 years.

2. Study area

Retezat Mountains is located in the western part of the Southern Carpathians as the first high massif (maximum altitudes of 2509 m asl in Peleaga Peak) facing the west wind masses (Fig. 1). It presents the marks of intense glacial erosion in the Pleistocene that left well-defined glacial cirques (Mîndrescu et al., 2014) and valleys (Urdea, 2000). Moraines from the Last Glacial Maximum are located down to about 1100 m asl on the northern and southern slopes (Urdea, 2000). However, repeated glacial cycles with a major cumulated effect manifested across the massif during the Quaternary period (Ruszkiczay-Rüdiger et al., 2016; 2021). The upper limit of the treeline is located at about 1800 m asl while above well-defined krummholz (subalpine vegetation level) and alpine levels are marked mainly by *Pinus mugo* and alpine meadows respectively. Other typical shrub vegetation consists of species like *Alnus viridis*, *Rhododendron ssp.*, *Juniperus sibirica*, *Vaccinium vitis-idaea*, *Vaccinium myrtifolium*, as well as the subalpine meadow communities. Glacial valleys have large and relatively flat valley bottoms with short and steep interfluvies. Several types of scree deposits like talus slopes, rock glaciers, debris-covered slopes, and block streams occupy a large part of the high altitude area of the massif (Onaca et al., 2017; Şerban et al., 2019) and are the result of frost weathering (Vasile et al., 2014; Vasile and Vespremeanu-Stroe, 2017). The dominant lithology is represented by granodiorite rocks but also by amphibolites and amphibolitic gnei-

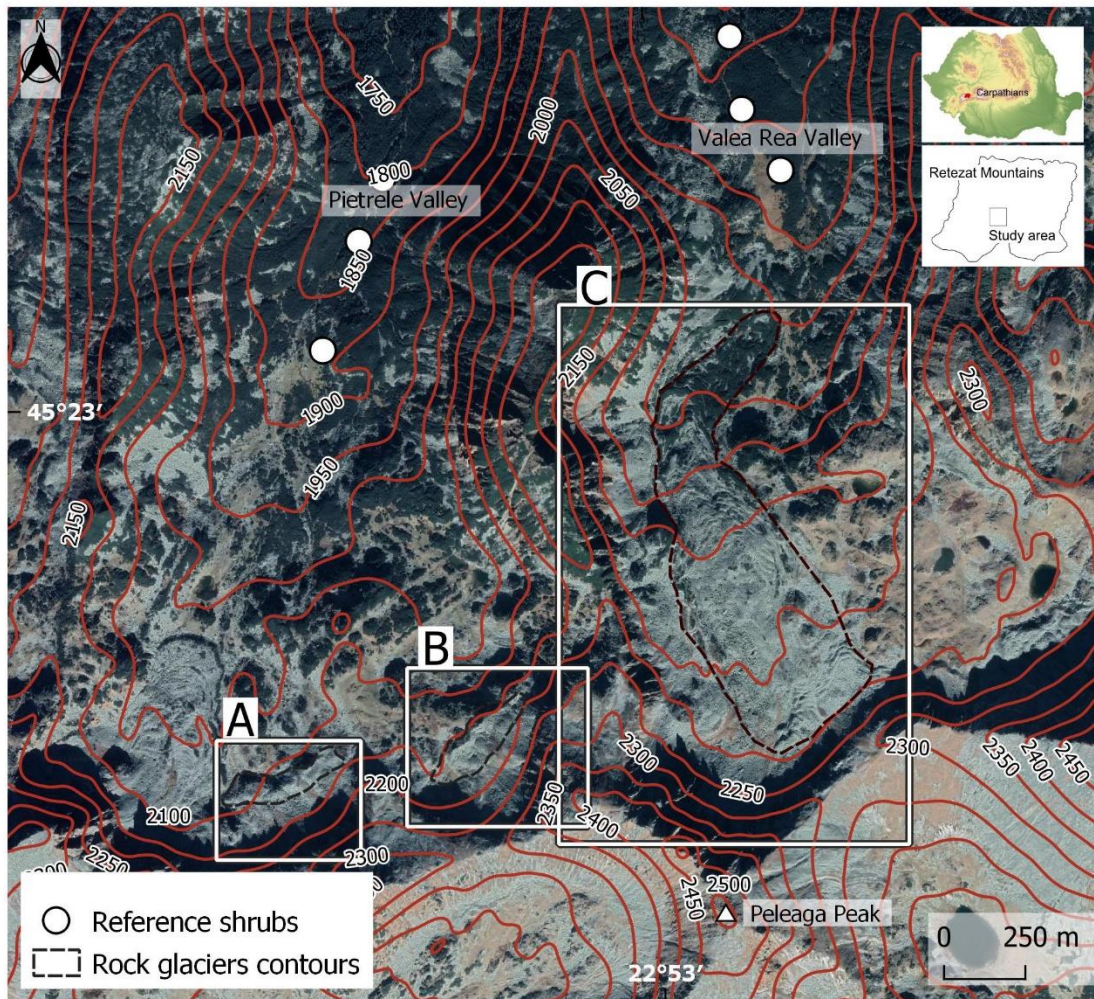


Figure 1 Study area in the central part of Retezat massif with Pietrele 2 (A), Pietricele (B), and Valea Rea (C) rock glaciers delineated. The shrubs sampled as reference are also visible here

sses in the SE of the massif in the Custura peak area, all of upper Precambrian age (Berza et al., 1989).

The climate is alpine specific to mountains located in mid-latitude temperate zones. The mean annual air temperature at the closest high altitude meteorological station (Vârful Țarcu, 2180 m asl) is -0.5 °C for the 1961-2007 time interval (Onaca, 2017). The 0 °C isotherm of MAAT considered for the Southern Carpathians occurs at approximately 2050 m asl, (Onaca et al., 2017) and the precipitation above 2000 m is 978-1330 mm/year (Micu et al., 2015). Most precipitations come as rainstorms during summer, and snow is generally present from November to June. Several patches of snow can persist until August in the most favorable locations, but perennial snow is not documented in Retezat. Sporadic permafrost is

found above 1950 m asl mainly in debris deposits like rock glaciers and talus slopes (Vespremeanu-Stroe et al., 2012; Onaca et al., 2015; Popescu et al., 2017). A large part of the Retezat Massif is included from 1935 in the Retezat National Park, which is the oldest in Romania, and it was UNESCO biosphere reserve from 1979 until 2021. The Retezat Mountains comprise a third of Romania's biodiversity but seriously affected by overgrazing in the last decades (Kiss and Alexa, 2015). The Carpathian range is a biodiversity hotspot in Europe serving as a palaeorefugium for numerous continental, arctic and alpine taxa during the Pleistocene glaciation because of the relatively reduced glacial cover in comparison to other mountain ranges (Bálint et al., 2011; Pușcaș et al., 2021).

The three rock glaciers investigated by dendrochronologic approach are located in Pietrele and Valea Rea Valleys (Fig. 2). Pietricele and Pietrele 2 are relatively small talus derived lobate rock glaciers. At the same time, Valea Rea is the largest rock glacier of the Romanian Carpathians, with a total length of more than 1.3 km, a surface of 0.41 km² and a complex origin and evolution (Fig. 3).

3. Methodology

3.1. Age determination of *Pinus mugo* shrubs

The age of *P. mugo* shrubs from Valea Rea and Pietrele valleys was investigated based on a dendrochronological approach. During field campaign conducted in 2013 on Pietrele Valley, 7 sites from Pietrele 2 rock glacier and 7 sites from Pietricele rock glacier were selected for sampling (Fig. 2). Each site is composed of 1 to 7 shrubs, the total number of individuals sampled is 36; they are located only on the terminal fronts of the rock glaciers. Additionally, 3 sites with a total of 11 individuals were collected as a reference in the Pietrele Valley in 2015 (Fig. 1). On Valea Rea valley, 19 sites (one shrub per site) along a longitudinal transect from the front upwards on Valea Rea rock glacier and an additional 7 reference sites (one shrub per site) were also selected for sampling in 2015.

Regarding ages, all values are reported to the year 2012. The main stem of each *P. mugo* individual was sampled either by taking cores (two cores for each stem sampled) using a Pressler borer or by taking discs with a hand saw. Disks or cores were labelled and their sampling location was recorded with a GPS device. In each site, the larger individuals were considered the oldest ones intended to be sampled. In the laboratory, samples were air-dried and then sanded using a sanding machine with sanding belts of progressively finer grits (60, 180, 220 or 320, 600), until growth ring structures became clearly visible. Standard dendrochronological procedures for age determination were applied (Stoffel and Bollschweiler, 2008; Stoffel and Corona, 2014). Because the age determination technique by simple ring counting allows only to estimate roughly an individual's age, ring-width measurements and cross-dating of growth-ring series are needed when

the dating accuracy is required. Therefore, after preliminary age estimation by ring counting, ring widths were measured to the nearest 0.01 mm using a LINTAB 5 measurement station (a measurement table coupled to a stereomicroscope Leica DMS 1000) and TSAP-Win Scientific software (RINNTECH, 2021). The quality of cross-dating and measurement accuracy of ring-width series was assessed visually and statistically using COFECHA software (Holmes, 1983), enabling the detection of missing and/or false rings. However, the ages obtained are possibly slightly underestimated because the growth rings of *P. mugo* may not be continuous along the stem, and as the shrub grows, the rings tend to build up only towards the terminal part of the stem.

3.2. Assessment of the rock glaciers vegetation cover and its changes

We calculated the actual vegetation cover, including shrubs and herbaceous vegetation, using supervised classification on each rock glacier in ENVI software on RGB orthoimages of 0.5 m spatial resolution from National Agency for Cadastre and Land Registration (ANCP), collection 2012. One of the major sources of error is related to the steep frontal slopes that present shades and could sometimes be classified erroneously as vegetation. To overcome this, we used newer alternative aerial images that did not have this problem.

Diachronic analysis of vegetation changes is done in a qualitative manner by using old aerial photographs obtained by the National Center of Cartography, formerly called IGFCOT in the 1968-1980 period (Mărgărint, 2010, p. 59). These images are not orthorectified and georeferenced so conclusions on the surface changes were made only on very small areas where deformations are negligible. Other old photos from the ground found in the literature were also used.

4. Results

4.1. *Pinus mugo* growth rings analysis

In Pietrele 2 rock glaciers front, the samples are located at altitudes between 2073 and 2118 m asl, and the oldest age from all the sites was obtained at S1, 182 years. The oldest age from each cluster of samples is commented. Two shrubs are older than

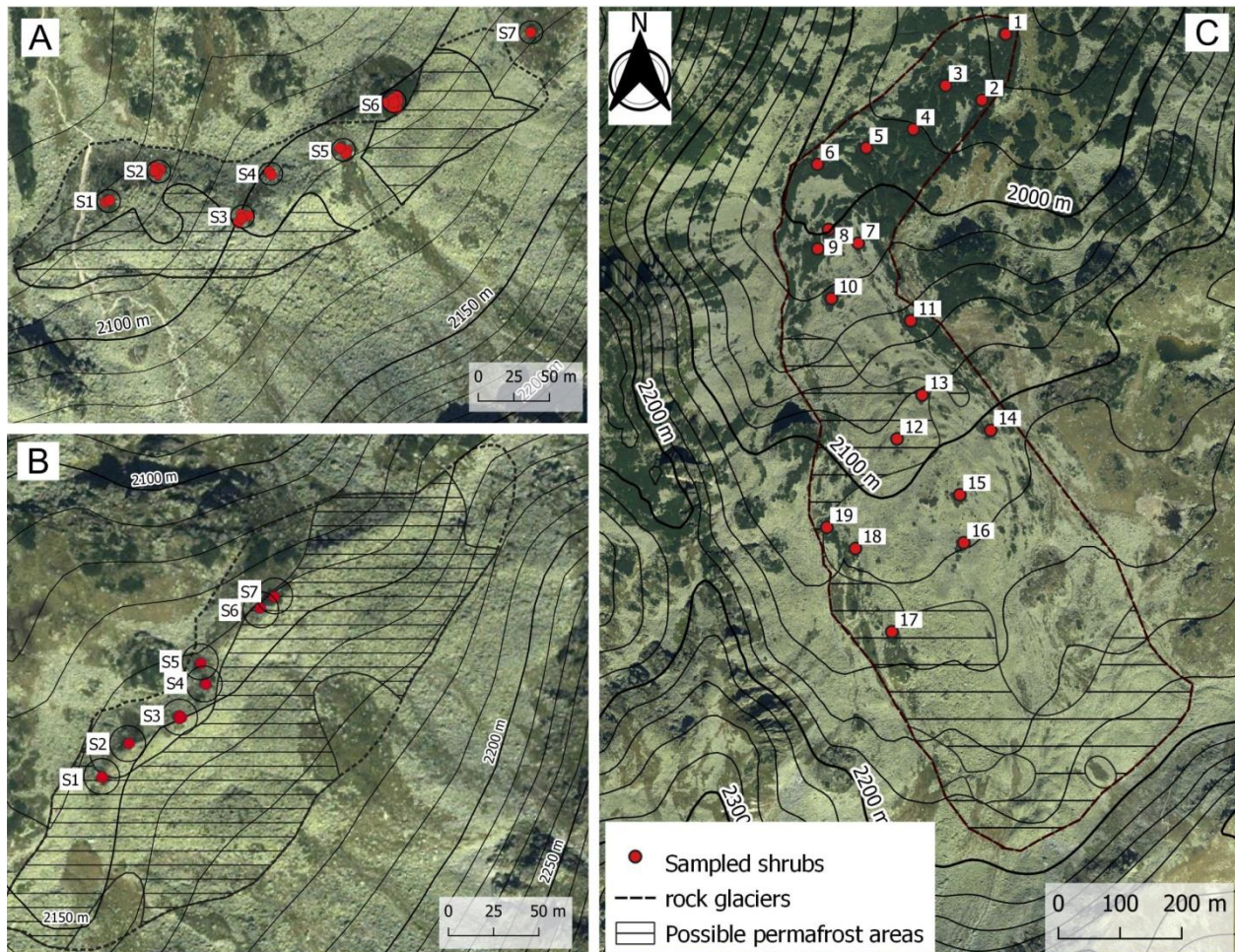


Figure 2 Location of sampled shrubs. In Pietrele 2 (A) and Pietricele (B) rock glaciers *P. mugo* shrubs were only located on rock glaciers fronts. In Valea Rea rock glacier (C) samples were collected on a transect from the lower to the upper part. The possible permafrost areas are determined by BTS measurements in 2011 (A and B) and 2013 (C)

150 years, five are older than 100 years, and six are older than 98 years (Figs. 4 and 5A). Only the one outside the rock glacier area is very young, of only 14 years. The shrubs and the sampling sites are distributed across the entire rock glacier front. In Pietricele rock glacier front, the samples are located in 2132-2150 m asl altitudinal interval, and one sample is older than 150 years, two are older than 100 years, and five out of seven are older than 50 years. The shrubs are only located on 2/3 of the entire rock glacier front length, while the remaining 1/3 is shrubs free (Figs. 2B și 3B). The distribution of values indicates a moderate tendency of shrub age decrease towards the northeast (Fig. 5B). In the Valea Rea rock glacier, the pattern of shrubs vegetation is represented by large cover in the lower terminal sector (2000-1950 m), with relatively low shrubs

vegetation mostly on ridges (1980-2140) only herbaceous vegetation above. The shrubs sampled are located between 1955 and 2133 m asl. The results indicate that two shrubs are older than 150 years, 11 are older than 100 years, 18 are older than 50 years, and only two are younger than 50 years. There is no spatial tendency of the age distribution. Comparing the entire data recovered from rock glaciers, age clustering at around 100 years with a larger number of cases towards higher ages can be observed (Fig. 5C). The reference shrubs are located at altitudes between 1790 and 1885 m asl and have, in general, younger ages in comparison to rock glaciers. They are clustered between 30-50 years old, but several shrubs are also old and comparable to those found on rock glaciers (Fig. 5D). The mean annual growth rate of shrubs located on the Valea Rea rock glacier

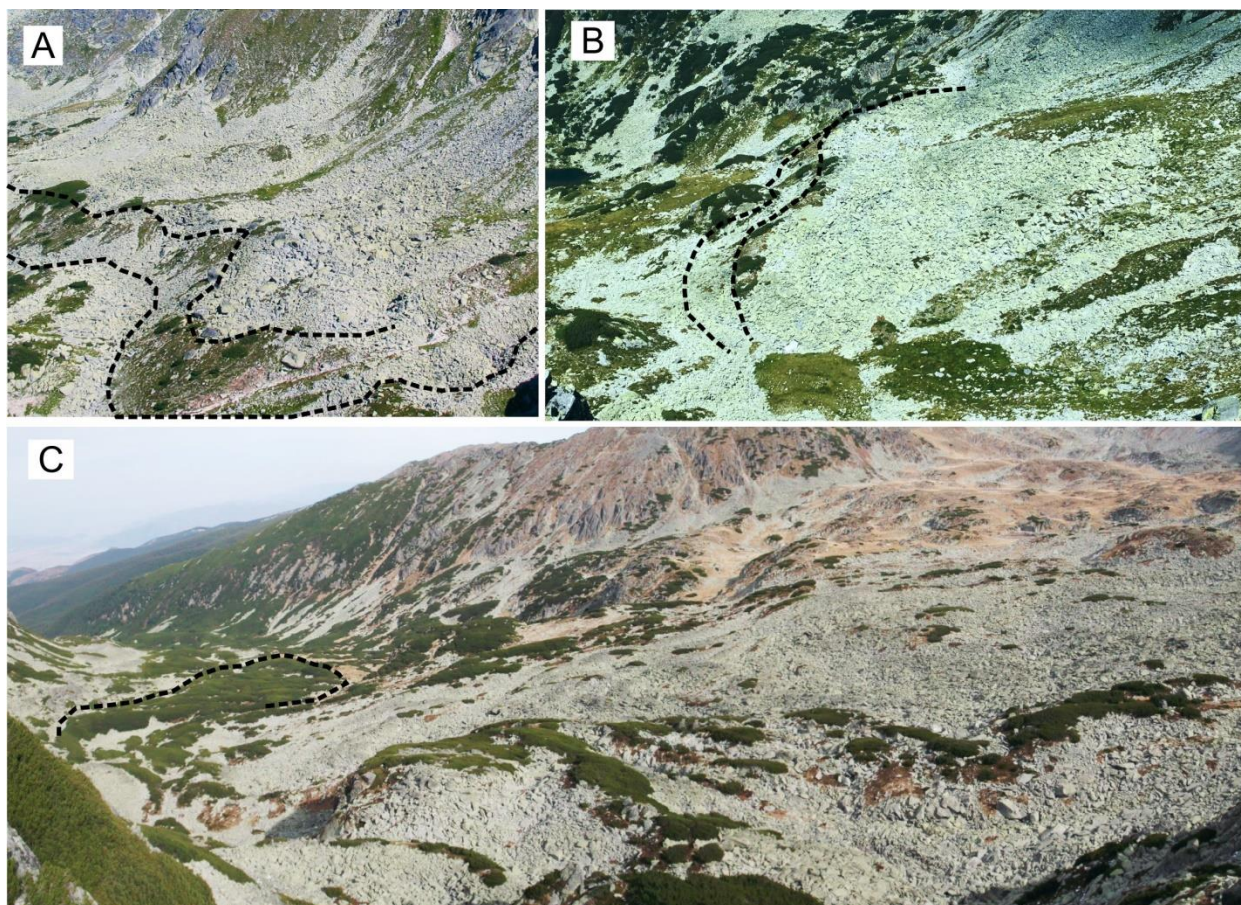


Figure 3 The three rock glaciers investigated by dendrochronological analysis with the terminal fronts delineated. Pietrele 2 (A), Pietricele (B) and Valea Rea (C- photo: Daniela Pascal, 2017)

is between 137 and 705 $\mu\text{m}/\text{year}$, and on the reference sites, it is between 205 and 1250 $\mu\text{m}/\text{year}$. There is a moderate to high correlation between growth rates and ages, with the older shrubs having lower values than the younger ones (Fig. 6). 63% of growth rates variability is related to shrubs' age.

4.2 Rock glaciers vegetation cover and its recent changes in Retezat Massif

An inventory comprising 82 rock glaciers from Retezat was used to analyze the vegetation cover and type visually. Six rock glaciers classes were distinguished (Fig. 7) and are described below.

1) Type 1 (e.g., Lower Ciufu Mic) rock glaciers are mostly covered with shrubs, but also trees can be found on their surface (16 rock glaciers);

2) Type 2 (e.g., Lower Ana) rock glaciers have densely shrub/grass-covered fronts and moderate shrub/grass cover on the rest of their surface (29 rock glaciers);

3) Type 3 (e.g., Pietrele) rock glaciers have (quasi)continuous shrub-covered fronts and sparse shrubs/grass patches on the rest of their surfaces (13 rock glaciers);

4) Type 4 (e.g., Judele) rock glaciers present sparse grass- and/or shrubs on their fronts, absent or very sporadic shrubs/grass on the rest of their surface are possible (11 rock glaciers);

5) Type 5 (e.g., Știrbu 2) rock glaciers with continuous grass-covered fronts, no shrubs, and insignificant grass areas on the rest of their surfaces (6 rock glaciers);

6) Type 6 (e.g., Galeșu) rock glaciers with sparse to absent grass on front and sparse to absent grass on the rest of their surfaces (7 rock glaciers).

The large majority of rock glaciers present *P. mugo* shrubs on their surface (the lower ones on almost their entire surface) shrubs coverage decreasing with altitude. Rock glaciers vegetation cover variation can be explained by the altitude chan-

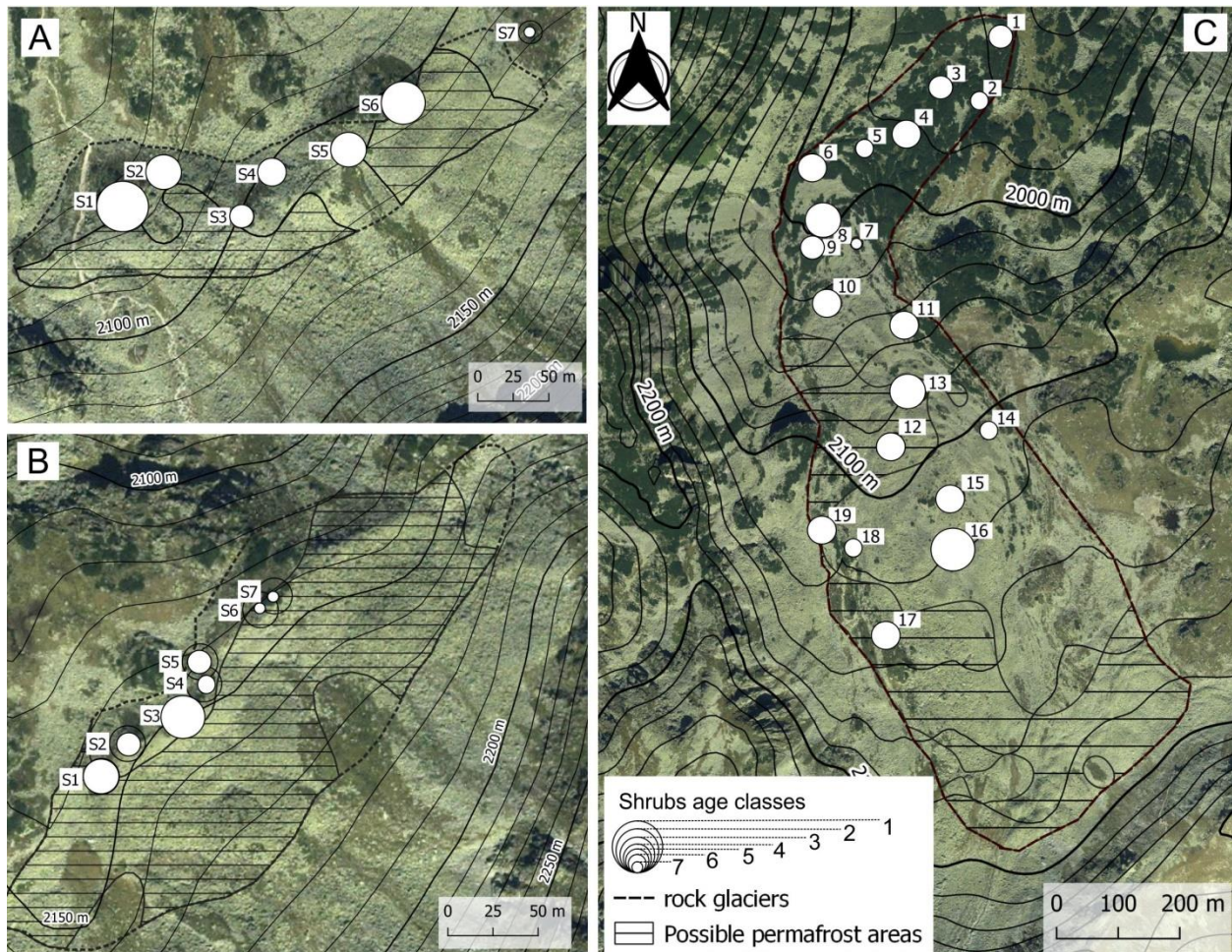


Figure 4 Shrubs age distribution at the studied rock glaciers. A - Pietrele 2; B - Pietricele; C - Valea Rea. Age classes in years are: 1 - >175; 2 - 151-175; 3 - 126-150; 4 - 101-125; 5 - 76-100; 6 - 51-75; 7 - <50

ge in proportion of 59% as revealed by the analysis of 40 rock glaciers (Fig. 8), simultaneously with MAAT decrease, snow cover persistence and wind intensity increase. The vegetation cover of more than 70% of the rock glacier surface can be found close to 1700 m asl, while less than 5% is encountered above 2150 m asl. The data are pretty scattered from the tendency line, especially at lower altitudes. For example, one rock glacier from about 1890 m asl has less than 28% vegetation cover, and another rock glacier as high as 2150 has nearly 80%. Other controlling factors that account for the rest 41% should be related to exposition that influences the input of solar radiation, debris size properties and availability of fines that allow plants to take roots but also the geomorphological processes that impede vegetation growth: snow avalanches, debris flows, rockfalls and debris dynamics (via rock glacier creep

and or dry creep). Quantifying the shrubs cover from krummholz is possible (Müllerová, 2005) as they have a higher contrast with the boulder mantle of rock glaciers from Retezat, while smaller shrubs and herbaceous vegetation areas need to be assessed with higher resolution images with better contrast.

The analysis of vegetation changes using old aerial imagery in the period between 1968-1980 and 2012 revealed that patches of *P. mugo* are relatively easy to distinguish from other land cover types, while grass cannot be separated from scree in most cases. Several rock glaciers from 1720 to 2130 m asl and from 1960 and 2180 m asl on the northern/southern slopes are presented (Figs. 9 and 10). In some cases, no identifiable modifications occurred (Figs. 9B, D; and 10B), while in other cases, the shrub areas increased their vegetation density but not their surface (Figs. 9A, C, and 10A, C). Only in one case stu-

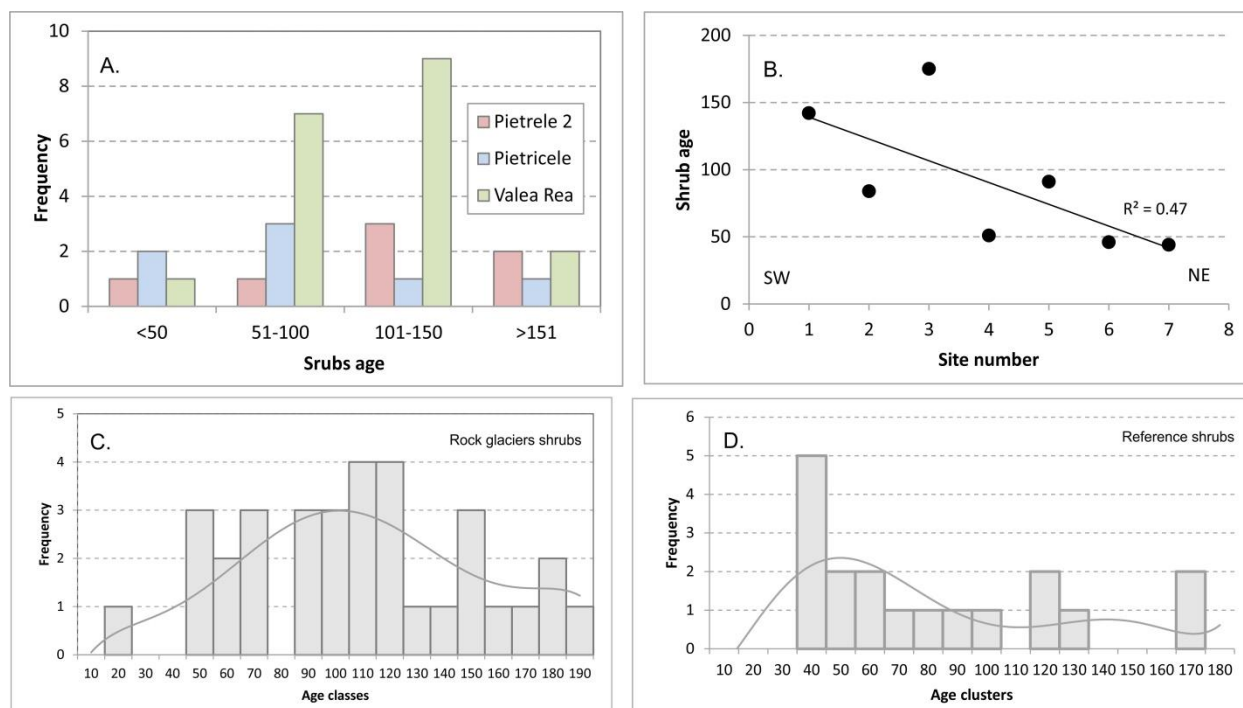


Figure 5 Statistic distribution of shrub ages on each rock glacier (A); the shrub age variation on the Pietricele rock glacier front from south-west to north-east (B); the absolute frequency of shrub ages on all rock glaciers classes (C) and on the reference sites (D); on the last two plots, the polynomial curve of 6th grade is added

dy was distinguishable the increase of shrubs area by the growth of new *P. mugo* small islands (Fig. 10A). In another place, we identified the first appearance of two trees within a compact shrub cover of *P. mugo* downhill the rock glacier front (Fig. 9A). For the Pietrele rock glacier, the current vegetation surface was compared with the one from more than 100 years ago using an old picture taken from Bucura pass (de Martonne, 1907). Across the rock glacier, the vegetation patches are at the same positions, but their extent seems smaller than in the present (Fig. 11). This is also visible at the rock glacier from 1950 m asl with the eastward exposition. In the background, on the steep slopes below the Stânișoara and Retezat peaks both the shrubs density and area seem to have increased considerably.

5. Discussion

5.1. Rock glaciers vegetation in Retezat Massif

Vegetation can be used as an indicator for rock glaciers activity and permafrost occurrence (Ikeda and Matsuoka, 2002). It is a good marker also for the activity of periglacial processes in high-altitude

mountains (Nedelea et al., 2012; Munteanu et al., 2012; Săndulache et al., 2015). Based on previous investigations in Retezat Massif, at type 1 rock glaciers permafrost is improbable, at type 2 permafrost is possible only at the rock glacier-talus slope contact but it is more a consequence of ground air circulation inside the talus slope, at type 3 rock glaciers permafrost is probable only at certain areas; at type 4-6 rock glaciers, permafrost is probable on almost the entire surface excluding the grass island (Vespremeanu-Stroe et al., 2012; Onaca et al., 2015; Popescu et al., 2017). According to vegetation indices, the 6th type rock glaciers are the only ones where permafrost creep could be likely. Their structure represented by numerous creep lobes with prominent ridges indicates a relatively fresh morphology even though lichens colonize all the exposed blocks except those located in depressions where snow persists more than 9 months/year (Burga et al., 2004). Rock glaciers from types 1-2 can be considered totally relict; types 3-5 could be considered pseudorelict (Onaca et al., 2017), and type 6 could be inactive. The latter seems to be of a younger chronological phase and with a morphology

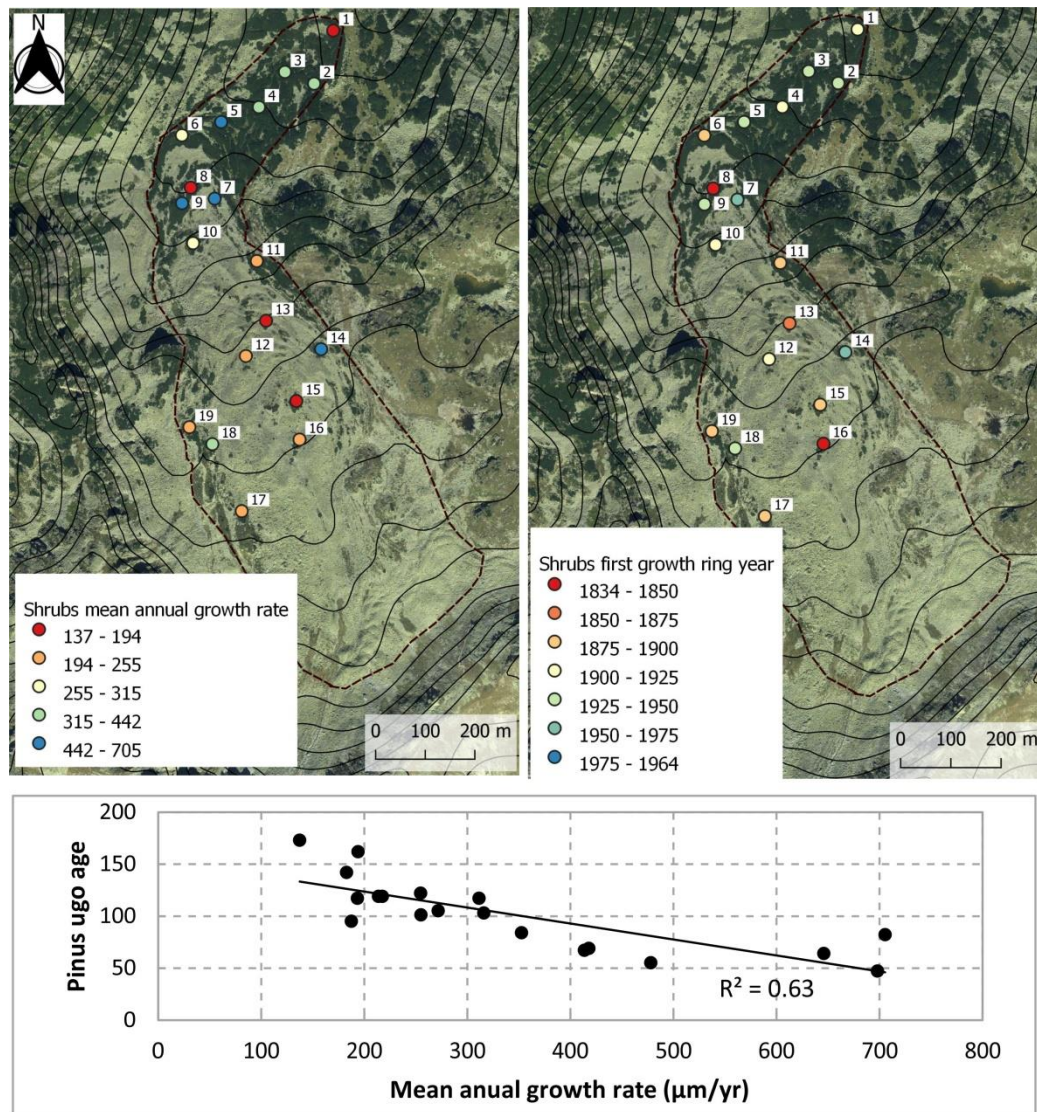


Figure 6 Mean annual growth rate and age of *P. mugo* from Valea Rea rock glacier along with the linear regression between the two parameters

that was probably more recently inactivated. Yet, age dating (relative and absolute) should be applied to distinguish between several generations of rock glaciers and verify this hypothesis.

5.2. Vegetation growth and changes in relation to rock glacier activity

Both Pietrele 2 and Pietricele rock glaciers belong to type 4 in terms of vegetation cover, presenting sporadic shrubs and grass on their frontal slopes and on the interior surface (Figs. 2, 3). Valea Rea rock glacier is a complex landform with progressive vegetation cover diminishing from its terminal front (1950 m asl) to its upper part (almost 2200 m asl) but

overall can be included in a type 3 rock glacier (Fig. 2).

Modern climate change causes the vegetation belts to increase in altitude and to decrease the intensity of periglacial processes that in turn allows a denser and wider vegetation cover. Rock glaciers are landforms restrictive to economic activities like grazing being less exposed to clearcutting. Thus vegetation colonization at these sites is impeded mainly by geomorphologic processes and climatic conditions. The oldest *P. mugo* shrubs from the three investigated rock glaciers date back to 1830-1839 and those from the reference areas date back to 1840 and 1888. The ages distribution indicates quasisynch-

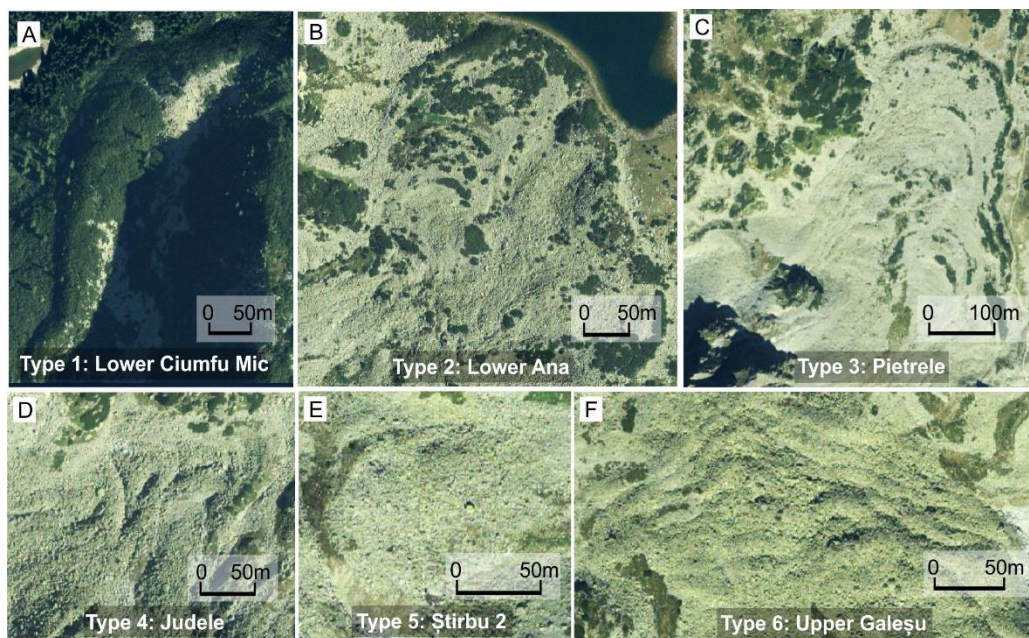


Figure 7 Rock glacier types in Retezat massif according to vegetation cover

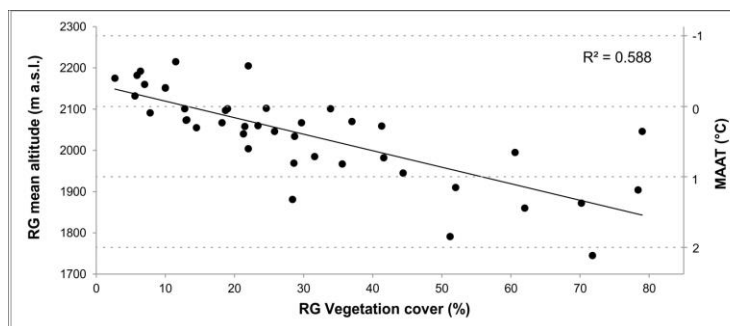


Figure 8 Rock glaciers vegetation cover, both shrubs and grass, in relation to their mean altitude

ronous colonization at Pietrele 2 and Valea Rea and relatively progressive colonization from SW to NE at Pietricele, where the northeastern most sector of the front is vegetation-free. Two scenarios can explain the shrub ages distribution and will be discussed below.

(1) In the climate influence scenario, the climate warming after the local Little Ice Age (LIA) allowed for the expansion of shrubs at higher altitudes starting with 1830 AD. A dendroclimatological study carried on in the Romanian Carpathians (Călimani Mountains) revealed that LIA last cold episode manifested between 1820-1840 AD after a long climate phase between 1370 and 1630 AD (Popa and Kern., 2009). In this case, rock glaciers were active in Late Glacial and probably early Holocene and stable

during most of Holocene, and only the moderate climate variations (e.g. LIA, Medieval Warm Period) could influence the *P. mugo* limit changes besides anthropic impact that is considered negligible on rock glaciers. The shrubs' age distribution pattern of the Valea Rea rock glacier supports this scenario.

(2) In the rock glacier movement scenario, the three rock glaciers were active in LIA and became inactive in the post-LIA period as a consequence of climate warming and subsequent permafrost degradation. That allowed the vegetation installation on the now stable rock glacier terminal fronts and internal creeping lobes. This is supported by the shrubs decreasing ages on Pietricele rock glacier front towards NE, but only if the northernmost sector would be active today, which doesn't seem to be the case considering the stable and weathered boulders

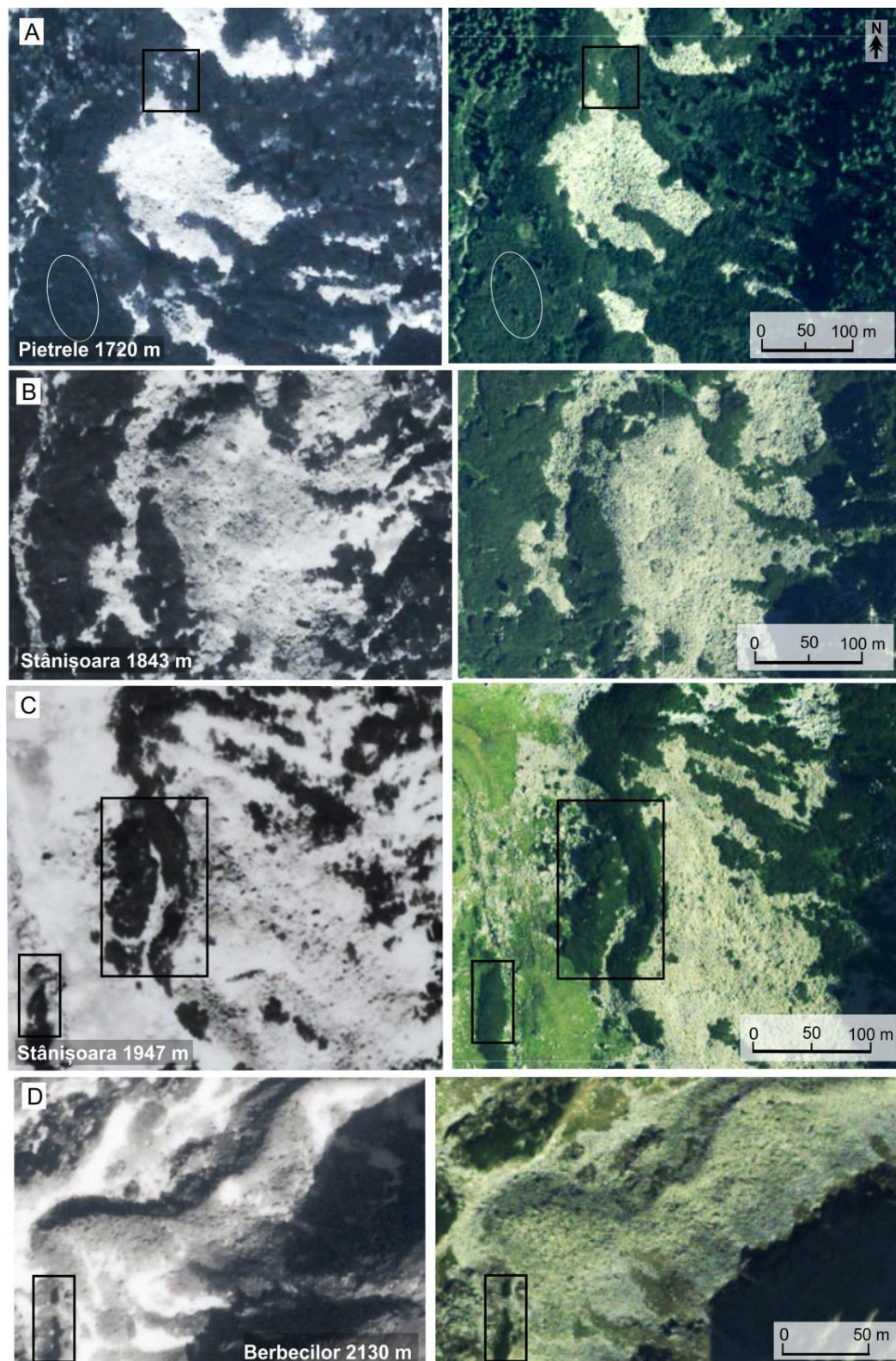


Figure 9 Vegetation changes on rock glaciers with northern aspects by comparing aerial imagery from 1968-1980 (left) and 2012 (right). The black rectangles indicate areas of increasing vegetation density. The old aerial imagery is the property of the National Center of Cartography (former IGFCOT) and was kindly provided by prof. M. Geanana (personal collection). The new images are the property of National Agency for Cadastre and Land Registration of Romania (ANCPI)

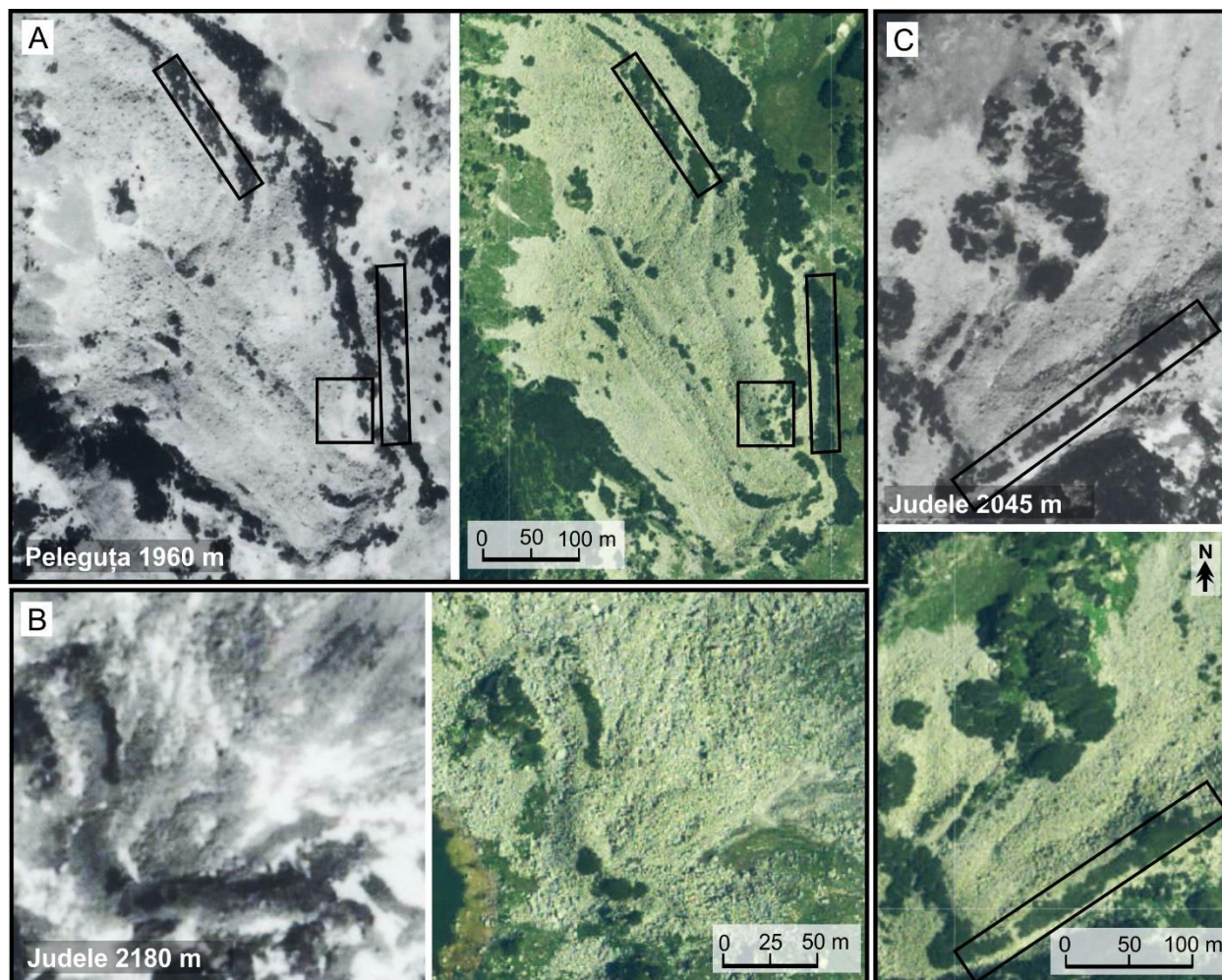


Figure 10 Vegetation changes on rock glaciers with southern aspects. The black rectangles indicate areas of increasing vegetation density and the appearance of new bushes (only on A). The old aerial imagery is the property of the National Center of Cartography (former IGFCOT) and were kindly provided by prof. M. Geanana (personal collection). The new images are the property of National Agency for Cadastre and Land Registration of Romania (ANCP)

even from this sector. This assumption is based on the fact that climate warming is immediate while the rock glacier dynamics decreasing response is delayed. This hypothesis was supported by Urdea (1998) that revealed a trend of decreasing ages of *P. mugo* from the main front inwards in the upper part of Pietrele rock glacier. In the south-eastern part of the rock glacier, the author indicates values from 137, 96, and 80 years old shrubs on three different set of ridges and 125 years in the central longitudinal ridge. That would correspond to 1860, 1901, and 1917 AD as years of shrubs development at these sites if we consider 1997 as the reference year. In Stânișoara Valley, the same author indicated the oldest shrubs developing at rock glacier/protalus rampart fronts

from 2130/2190 m asl in 1857 AD and 1895 AD, respectively interpreted similarly as a result of the progressive cessation of permafrost creep with altitude followed by subsequent shrubs colonization. If we assume an average lapse rate of $0.6\text{ }^{\circ}\text{C}/100\text{ m}$ and a $-0.5\text{ }^{\circ}\text{C}$ MAAT from Țarcu Mountain meteorological station to be valid for Retezat at 2180 m asl, in that case, the rock glaciers investigated in this study and those of Urdea (1998) would have a MAAT at their front of $-0.6\text{...}+0.9\text{ }^{\circ}\text{C}$. The necessary regional (average) MAAT for relict rock glaciers to be active was calculated to be $-1.6\text{...}-3.3\text{ }^{\circ}\text{C}$ in different mountain regions in Switzerland, even though local differences can be encountered (Frauenfelder and Kääb, 2000). Also, Haeberli (1985) indicated that acti-

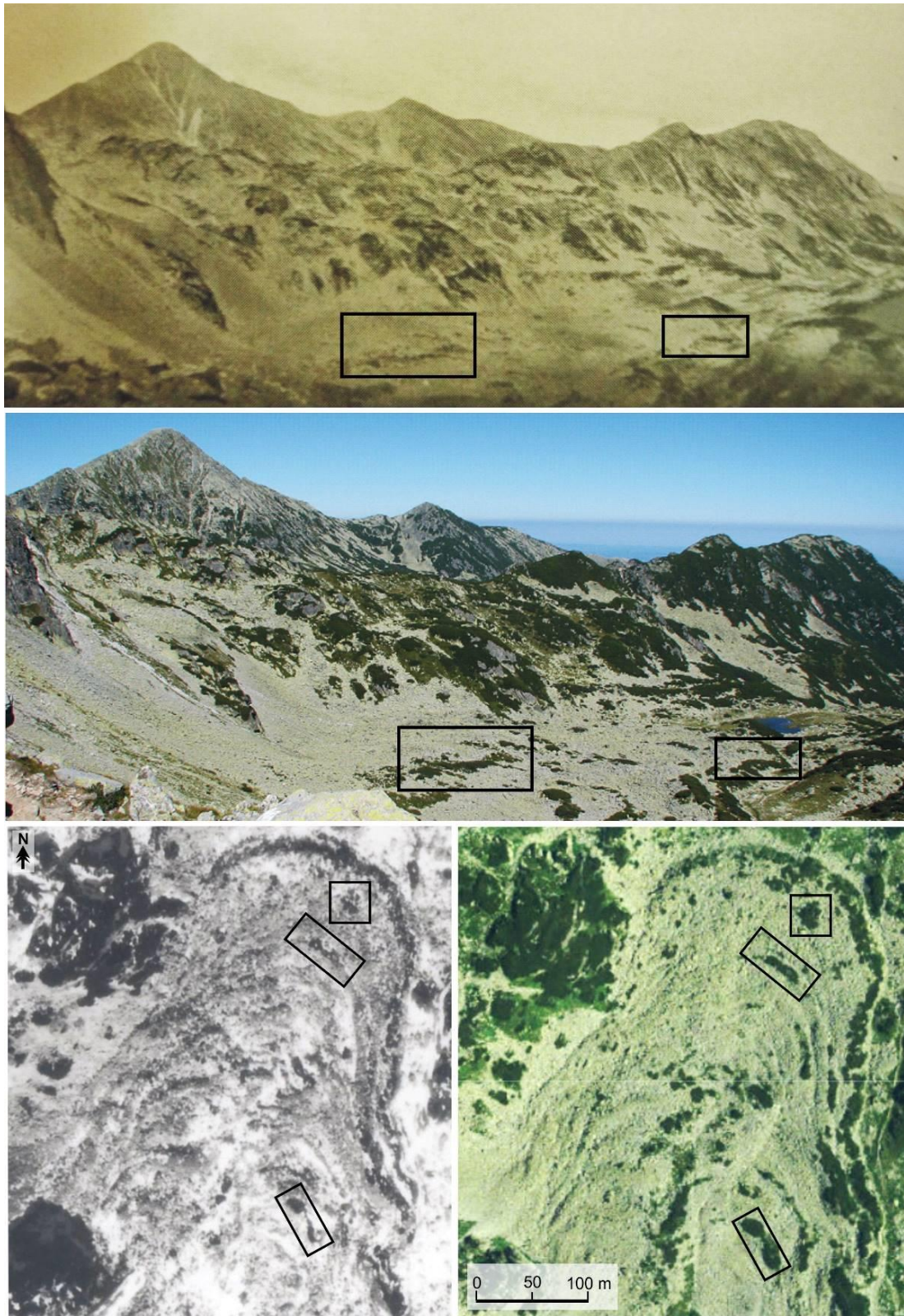


Figure 11 Vegetation changes on Pietrele rock glacier from 1907 to 2012 and from 1968-1980 to 2012. The black rectangles indicate areas of increasing vegetation density. The old picture of Pietrele rock glacier (up) is from de Martonne (1907)

ve rock glaciers typically lie above the MAAT isotherm of -2°C . Considering the relatively dry alpine climate of Southern Carpathians, it is expected that Retezat rock glaciers needed MAAT rather closer to -3°C . Thus, the temperature lowering of several degrees Celsius needed to allow rock glacier activity would then have been possible only in Late Glacial and Early Holocene (before 11.1 ka and during 8.2 ka cold event) as LIA climate conditions were insufficiently cold. However, Necşoiu et al. (2016) revealed that Pietrele rock glacier is active in the upper part and Judele rock glaciers is active on its entire surface with very low creep rate of a few cm per year. This is a much weaker dynamics, typical for marginal permafrost conditions, in comparison to what is traditionally considered an activity for a rock glacier, movement of a few tens of cm and m per year (e.g. Serrano et al., 2006) besides the recent rock glaciers acceleration caused by climate change (Kääb et al., 2021). In that case, shrubs growing seem possible in spite of small boulders movement. Suppose the permafrost creep was more intense in comparison to the recently discovered values as a consequence of colder LIA climate. In that case, it might be possible that the rock glacier movement scenario is valid for explaining the shrubs age distribution at the six investigated landforms. Recent studies indicate indirectly that climate was milder than at present in 8-3.2. ka BP as indicated by timberline at least 50 m higher than today at lake Lia in Retezat (Vincze et al., 2017). The fact that *P. mugo* ages are higher on rock glaciers despite of the more difficult conditions indicates the anthropic activity that manifested already beginning with 2.2 ka BP as indicated by the above mentioned study. This could also be the case for the lower, highly vegetated part of Valea Rea rock glacier. In any case, the Valea Rea shrubs age distribution presented in this study does not support a progressive upwards cessation of permafrost creep even though the conditions of this rock glacier are ideal (high altitudinal range of almost 250 m and numerous creep lobes systems that could move independently) for illustrating such a scenario. Based on the presented arguments, we incline more to the first scenario but the second scenario cannot be excluded. Future investigation like age dating of

landforms should give a clearer picture of the rock glaciers' past evolution.

The vegetation changes from 1907 to the present indicate the vegetation expansion of *P. mugo* on steep mountain slopes. However, on rock glacier ridges, the *P. mugo* was already in place more than a hundred years ago, but it increased its surfaces since. In the period between 1968-1980 and 2012, the rock glaciers shrubs seem to have increased the density and only sporadically it has colonized new debris areas across rock glaciers, which proves that vegetation colonization on rock glaciers was at its maximum immediately after LIA (1840-1880) or at the end of the 19th and first half of the 20th centuries. The tendency of *P. mugo* area growth is natural in the context of climate change even though in the Carpathians the clearcutting surpasses the shrubs' natural expansion (Roşca et al., 2019). In this context rock glaciers shrubs seem to be more useful in climate reconstruction than shrubs from surrounding areas.

6. Conclusions

In this study, rock glacier shrubs vegetation cover, the timing of colonization, and vegetation cover changes from the last century were investigated in Retezat Mountains. The following conclusions could be drawn:

- At all the three investigated rock glaciers, the earliest period of *Pinus mugo* colonization correspond to the fourth decade of the 19th century, matching probably the end of the last cold spell of Little Ice Age;
- The frequency of very old shrubs (>150 years) is much higher on rock glaciers in comparison to the reference areas indicating a probable less anthropic to non-existent impact on the former;
- Rock glaciers' vegetation cover increases with altitude, and it has changed slightly after 1907 AD, mainly by increasing vegetation density and slightly increasing the surfaces of existing patches. New islands of shrubs appeared only rarely on rock glaciers. Six types of rock glaciers were classified based on their vegetation cover that has relevance for potential permafrost content evaluation and activity status inferences;

- It is still debatable whether vegetation colonization of rock glaciers was caused by post LIA climate warming or by rock glacier inactivation. Our results indicate rather the first scenario as more probable as the vegetation growth was relatively

synchronous on the largest rock glacier from Romanian Carpathians. That invalidates the possibility of progressive inactivation of several creeping lobes that should have caused the gradual vegetation settling.

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A history of the circum-Pontic river channels marked by climate and sea level changes during the Late Quaternary (25-8 ka BP)

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ABSTRACT

The present paper synthesizes the evolution of fluvial systems tributary to the Black Sea (within 25 – 8 ka BP time frame) by reviewing the existing paleoclimate, hydro-geomorphology, sedimentology, and Late Quaternary chronology. These studies are unevenly distributed throughout the Black Sea (BS) drainage basin and were used to decipher the river response to climatic, eustatic, tectonically induced or vegetation cover changes. The river response was sensitive overall and is visible particularly in terms of the channel planform. The alternation of cold and warm climate phases has controlled to a significant degree the water and sediment inputs along the drainage networks, such that river channels planforms shifted from the braided type to the anastomosed type (with various transitional phases, either wandering or anastomosed), and from macro-meanders to small scale meanders. The climatic model of fluvial evolution was altered by the other controlling factors specific to the large BS drainage basin, including tectonics (in the Pannonian Plain, Western Plain of Romania, Lower Siret Plain), the proximity to coarse-grained sediment sources (typical for medium-sized Carpathian tributaries), the obstruction exerted by glacial sediments generated by the ice cap retreat (as illustrated on the Upper Dnieper), eustatic oscillations (detected on the Danube or Sakarya River). The sea-level drop by more than 100 m during the Last Glacial Maximum resulted in the conversion of ca. 30% of the present sea surface to dry land (mostly on the NW continental shelf); several sinuous-anastomosed paleo-channels pertaining to major rivers (Danube, Dniester, Dnieper) with depths ranging between 30 and 90 m were reconstructed on the surface of this emerged land. Black Sea became connected to the Planetary Ocean during the Early Holocene (ca. 9.4 ka BP). The sea waters flooded extensive areas at the NW and advanced upstream along tributary river valleys. The effects of base-level rise were reflected in the anastomosed style of river channels and the accumulation of deltaic formations.

KEYWORDS fluvial style reconstruction, paleo-hydrological changes, base-level oscillation, Black Sea drainage basin, Late Pleniglacial, Late Glacial

1. Introduction

For more than two decades the Black Sea has been at the core of a heated scientific dispute regarding major level oscillations in the Upper Pleistocene and the Holocene. The most compelling argument is related to the so-called „flood question“ (originating from the Noah flood myth and expressed in the title of the books published by William Ryan and Pittman, 1998 and Yanko-Hombach et al., 2007). Based on an investigation of marine geology widely regarded as remarkable (apud Baker, 2007), Ryan et al. (1997, revised in 2003) hypothesized that the Black Sea was abruptly flooded during the Early Holocene (at ~8.4 ka) and linked the event with the Biblical myth of the Flood. The scientific community was surprised by this interpretation, while the hypothesis had captured the interest of both the media and the general public. In short, this theory assumes that during the last Quaternary glaciation, the Black Sea was a freshwater lake whose level was as low as -120 m (up to -140 m) below the present level. At ~8.4 ka the waters of the Mediterranean Sea swiftly pierced through the Bosphorus strait such that in a very short time frame of approx. 2 years the basin was filled and the water salinity increased (Lericolais, 2017). The sea-level rise resulted in the flooding of more than 100,000 km² of land which is currently part of the continental shelf. According to the authors, the abrupt flooding event led to the sudden migration of populations from around the Black Sea to the inner regions of Europe. Moreover, the authors accredit the idea that the legend of the Biblical Flood originated in this catastrophic event which took place in the Black Sea region instead of the Biblical Mesopotamia.

This bold hypothesis proved to be rather divisive in the scientific community, generating both support and skepticism. However, the rising interest resulted in several investigations undertaken by international expeditions using advanced research equipment which yielded numerous publications on the matter.

Although research has intensified in the Black Sea region, the scientific community is still navigating, to a large extent, a realm of hypotheses, which have gradually multiplied to three main concepts (i.e., catastrophic - Ryan et al., 2003; gradual - Hiscott et al., 2007; oscillating - Chepalyga, 2007),

and later on to four (cf. Yanchilina et al., 2017), as the amount of evidence has augmented as well as the complexity of data interpretation.

Some of the questions persisting on the topic were summed up as paradoxes (Yanko-Hombach and Kislav, 2018) which originate in several anomalies, including the ones related to the major oscillations of Caspian and Black Sea levels (which were connected repeatedly throughout the Quaternary), to the water sources which generated the oscillations, to the insufficient available numerical data and the low correlation with climate events, or to some contradictions between various datasets yielded by scientific investigations. Among these diverse opinions, however, one theory has gained broad support, stating that the most recent connection of the Black Sea with the Mediterranean Sea occurred during the Early Holocene between 7.5 ka and 9.4 ka, more specifically at 8995 ± 145 cal a. BP (cf Soulet et al., 2011, which reviews state-of-the-art analysis and calibration techniques).

This work does not attempt to discuss in detail all the arguments and evidence supporting either hypothesis (some important references are also included in the bibliography of this paper) or assert opinions regarding the previously mentioned debate. Instead, we focused our attention on a category of landform ensembles that are sensitive enough to preserve the traces of major climatic and eustatic shifts of the Late Pleniglacial as part of their sediment and morphological archives (Vandenberghe, 2002, 2003). River valleys are such landform ensembles suited for this type of approach, more specifically the valleys or rivers tributary to the Black Sea or circum-Pontic rivers, in our case. Danube, Dniester, Dnieper, Don, to name just the largest circum-Pontic watercourses, were forced to adjust their morphological parameters in time such that the geomorphic work can be accomplished: transferring extremely variable amounts of water and sediment to a highly oscillating base level.

Consequently, a significant number of studies of paleoclimatology, hydro-geomorphology and Late Quaternary chronology regarding the rivers of the circum-Pontic region were reviewed. Considering the broad geographical distribution of areas where

research data is available within the region of the Black Sea basin, the studies were grouped into three categories:

(a) paleo-geomorphological reconstructions based on fluvial sediment archives (particularly for Danube, Dniester and Dnieper, as well as for two major Anatolian rivers, Kızılırmak and Sakarya);

(b) paleo-hydrological reconstructions based on the mathematic modelling of paleo-channel morphology (as applied on the fluvial systems of the Russian Plain, such as Bug, Dnieper, Don and their tributaries);

(c) reconstructions of Black Sea level oscillations and effects on paleo-drainage on the continental shelf. The results of reviewed studies were discussed against the background provided by various climate variability records (following the recommendations of INTegration of Ice-core, MARine, and TERrestrial records INTIMATE community, Blockley et al., 2012), as well as by other regional interpretations regarding the response of fluvial ensembles to the combined action of control factors in LQ.

2. Regional setting

The drainage basin of the Black Sea (BS) extends over nearly 1.9 million km² (without Black Sea surface area) (Vespremeanu and Golumbeanu, 2018), from Central Europe (west) to the Russian Plain (north), the Caucasus (east) and the Pontian Mountains and the Anatolian Plateau (south) (Fig. 1). Whereas the Danube is the largest single tributary of BS, with a drainage basin of ca. 801,000 km², the rivers draining the Russian Plain cover more than 1 million km² (Table 1). The sources of the water discharge transported to BS originate in the Alps (partially), the Carpathians (tributaries of Danube and Dniester) and the Valdai Plateau (Dnieper and Don). The southern tributaries of BS are typically shorter in length with comparatively smaller drainage basins. Of these, Kızılırmak (Red River) and Sakarya, which originate in Central Anatolia, are more notable. For the purpose of this investigation, only the rivers included in Table 1 are discussed, which were selected based on the availability of paleo-evolution studies in their morphological ensembles.

The Black Sea is a semi-closed Mediterranean basin with a marginal position: as deep as 2250 m

and over 466,200 km² in area (Vespremeanu and Golumbeanu, 2018). Due to its position of semi-isolation within the planetary ocean, the Black Sea tends to amplify environmental changes, and thus its detailed and sensitive paleoclimatic record has become a focus of oceanographic research (Yanko-Hombach et al., 2007). During its Quaternary history, the BS has undergone numerous stages consisting of alternating phases of complete isolation from the ocean and connection with the Caspian Sea through the Manych channel, ensued by reconnection with the Mediterranean Sea through the Bosphorus (Fig. 1). Throughout this time frame, the BS level has oscillated by as much as 120 m while undergoing large variations in terms of water salinity and a vast array of derived ecosystem shifts (Lericolais, 2017).

As regards the climate of the study area, the drainage basin of the Black Sea falls under the following dominant types: (i) the central European, or transitional type of climate, resulting from the interaction of both maritime and continental air masses; (ii) the continental type of climate dominating a large area pertaining to the Eastern European Plain, to the north of the Black Sea; (iii) the continental type of climate, with humid subtropical influences in some sectors of the seafront (Crimea, Caucasus) and drier traits on the Anatolian seafront. The current types of climates were also characteristic for the Pleistocene environment in this region, albeit with a higher degree of continentalism (i.e., lower temperatures and scarce precipitation during cold seasons, linked to the advancement of the icecap) (Starkel, 1977; Frenzel et al., 1992).

In terms of morphotectonics and morpho-structure, the drainage basin of the Black Sea is characterized by the ensembles created by the Alpine orogenesis (the Alps, the Carpathians, the Caucasus) and the great Eastern European platform. To the south, the drainage basin of BS overlies the region dominated by the North-Anatolian fault, which separates the Eastern European Plate from the Anatolian Plate.

Major fluvial systems which shape this region are presented in Table 1. The availability and spatial coverage in terms of paleo-evolution studies are uneven, at best, both in terms of distribution along the river valleys and temporal coverage of the review.

Table 1 Data related to major tributaries of the Black Sea for which paleo-evolution studies are available and have been reviewed in this paper

River	Drainage basin area, (A, km ²)	River length (L, km)	Water discharge (Q, m ³ /s, annual mean)	Qmax, m ³ /s (mean maximum)	Suspended sediment load (kg/s)
Dnister	68,627	1362	310	610 (June 1969 = 3000)	550
Southern Bug	63,700	806	108	-	63.4
Dniپر	531,817	2145	1670	-	254
Don*	425,600	1870	935	-	-
Kuban*	58,000	870	425	-	-
Danube	801,463	2850	6515	21,867	1534
Tisa	156,087	966	792	-	-
Mureş	30,332	789	186	-	-
Siret	44,835	706	254	2643/4650	275
Kizirlimak **	78,646	1355	185	-	-
Sakarya **	56,504	824	177	-	120

* tributaries of Azov Sea

** major southern (Anatolian) tributaries of BS



Figure 1 Black Sea drainage basin with the most important tributaries (modified after <https://commons.wikimedia.org/wiki/>). Location of the paleo-fluvial and paleoclimatic study areas used in this article (Numbers next to the sites refer to the site names in Annex 1); red stars: sites with large paleochannel remnants (after Sidorchuk et al., 2011)

wed time frame (Pleniglacial, Late Glacial, Early Holocene). For this review we have focused on the Lower Danube and its Carpathian tributaries, the rivers draining the Eastern European Plain and the two most notable Anatolian rivers (Kızılırmak and Sakarya) (to illustrate, investigated reaches and sites where paleo-evolution data is available are pinpointed in Fig. 1 and described in Annex 1). Along with paleo-fluvial data we have used other paleo-environmental information available for the neighboring areas of BS, such as the stalagmite isotopic records obtained at Sofular Cave in north-western Turkey (Fleitmann et al., 2009) or at Karaca Cave in north-eastern Turkey (Rowe et al., 2012). Similarly, we have included in this review the records of Tăușoare Cave in the Eastern Romanian Carpathian (Staubwasser et al., 2018) and the lacustrine sediment pollen records at Preluca Țiganului, in NW Eastern Romanian Carpathian (Feurdean et al., 2014).

According to their nature, fluvial sediment archives typically have much lower resolution in recording the main paleo-environmental and climatic phases compared to other terrestrial records (such as stalagmites, cave ice cores, lakes and peat bogs etc.), as fluvial memory has mainly retained only major climatic, eustatic or tectonic shifts. The most accurate, high-resolution record of past climate for Early Quaternary is the isotopic $\delta^{18}\text{O}$ curve from Greenland (Dansgaard et al., 1993; Rasmussen et al., 2008; Svensson et al., 2008), which is used by the INTIMATE group as a template for reporting terrestrial records of the Northern Hemisphere (Moreno et al., 2014). A stratigraphic division of the last cold stage in Europe (Frenzel et al., 1992) distinguishes the following climate phases during the studied period:

(a) Upper Pleniglacial (28 - 14.65 ka cal BP, includes the LGM) - coldest and driest period but with several weak warming fluctuations;

(b) Late Glacial (14.65 - 11.65 ka cal BP) - was characterized by gradual warming with significant fluctuations (including the cold Younger Dryas: 12.7-11.65 ka BP) visible in the expansion of the forest belt from the east and melting of permafrost;

(c) Early Holocene (11.65 - 8.5 ka cal BP) - was characterized by rapid warming.

We used these major climate phases as time frames, according to which we integrated fluvial change records as reported by reviewed study cases (see Annex 1).

3. Paleo-geomorphological reconstructions based on fluvial sediment archives

Numerous studies on fluvial sediment archives have been published across the board, particularly in temperate environments, as mounting evidence showed that fluvial morphology and sedimentology have the capacity to store paleo-geomorphological information (Schumm, 1977; Starkel, 1991; Vandenberghe and Sidorciuk, 2020). Research works carried out on the tributaries of mid-course and Lower Danube, on the Danube itself, as well as on the largest rivers and the tributaries of the Russian Plain, illustrate how LQ climate changes have generated tangible, dramatic responses of river channels. Our intent was to collect as much data as possible, including substantial results, regarding the behaviour of rivers at contact with BS, which has acted as a base level for the geomorphic activity of tributaries. The scarcity of detailed LQ paleo-fluvial in this area has prompted us to turn our attention upstream to the valleys of these tributaries where such investigation results are available. Finally, we were interested to observe whether there was any synchronicity in the response of rivers to major climate shifts, if other factors intervened in the behaviour changes of river channels (such as tectonics and eustatism), whether this synchronicity was limited to certain morpho-structural units of the BS basin, or, conversely, if this was a dominant phenomenon for all river channels in the investigated region.

Relatively abundant research has been carried out on the paleo-geomorphology of rivers from the Pannonian Basin, particularly on Tisa and its tributaries, with hundreds of cores retrieved and dated using OSL and radiocarbon dating and palynological assemblages established based on successions of fine-grained sediments in order to determine the history and controls on alluvial sedimentation for throughout more than 30 ka (Gabris and Nador, 2007; Kasse et al., 2010; Nador et al., 2011; Şipos et al., 2016; Vandenberghe et al., 2018 and their references). By combining controls of fluvial

processes in a summarized illustration (Fig. 2), several river pattern change phases are distinguished, driven mainly by climate changes, as follows: (i) braided channel at 23 – 19 ka, overlapping the Last Glacial Maximum (LGM), and at 16.5 – 14.6 ka (Oldest Dryas), against a dry cold climatic backdrop, when Tisa and its tributaries received large amounts of sediments they were unable to transport and were thus stored as scroll bars and islands. Moreover, fluvial dynamics was complemented by intensive wind activity, noticeable in the landscape of this region due to the wide-presence of fluvial sands prone to wind

reworking during dry phases; (ii) meandering channel characteristic for warmer, more humid periods, when vegetation abundance resulted in increased dune stabilization and soil formation, thus reducing the sediment transfer to river channels. Prevalent processes include incision and shaping of fluvial terraces. The dominant pattern relied on the meandering, but meander size varied: e.g., meanders which formed between 19 and 16.5 ka had lengths and amplitudes 3 to 4 times greater than current meanders due to proportionally higher streamflow discharge values.

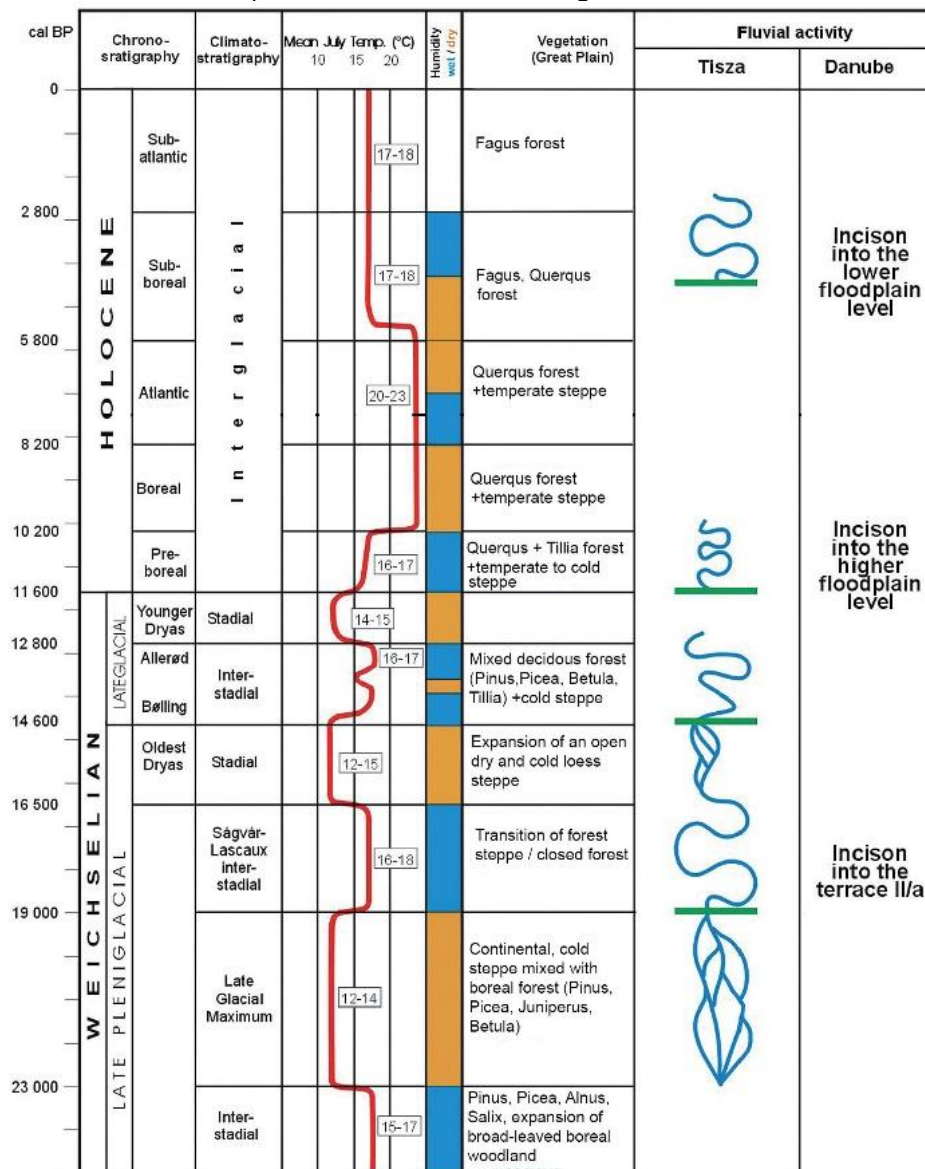


Figure 2 Paleo-environmental changes in the Pannonian Basin and fluvial activity of the Middle Tisza region and the Körös (Criș) basin during the last 25 ka (modified after Gabris et al., 2012)

Downstream of the Pannonian Plain, a group of rivers of interest include the tributaries of Danube, which drain the Carpathian arch. These are typically smaller-sized rivers, with drainage basins that seldom surpass

10,000 km² and their discharges vary in the range of tens of m³/s. The investigations on their LQ fluvial evolution were recently reviewed and summarized (Fig. 3).

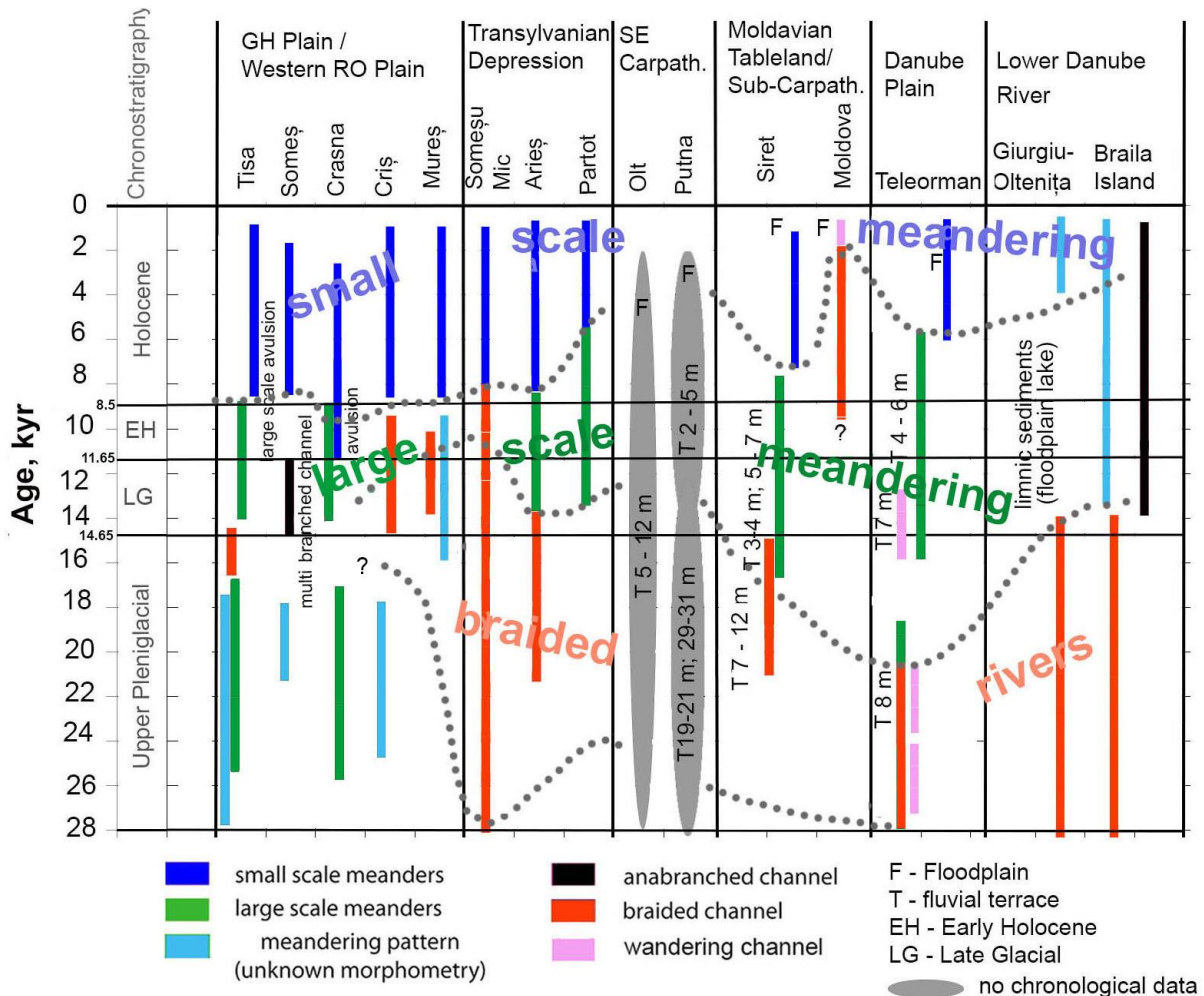


Figure 3 The ages of the lower fluvial features, as reflected by the recent studies on Late Quaternary fluvial evolution in several morpho-structural units from Romania (reinterpreted after Perșoiu et al., 2017, with additions)

Compared to rivers of the Pannonian Basin, whereby the bed material is predominantly fine-grained, the rivers of the Transylvanian Depression and the outer rim of the Carpathians are typically coarser-grained, with beds dominated by gravel, and are characterized by higher energy environments, due to greater slope gradients and shorter distances to coarse sediment sources. Consequently, the response of channel configuration was different in terms of the ability to generate mature channel styles. For instance, during a long period of time (Pleniglacial to Late Glacial) the rivers developed

gravel-bed braided channels with a tendency towards aggradation, albeit they evolved throughout distinct climatic phases. The first major change in fluvial morphology was documented during the Late Glacial when numerous rivers abandoned the braided style for large-scale meandering rivers (i.e., meander size was 3 to 5 times greater compared to present-day meanders). The second major shift occurred in the Early Holocene when rivers maintained their meandering styles, although the meander size decreased significantly. In the case of Carpathian rivers, some synchronicity is evident in terms of the

fluvial activity, which is regarded by Starkel et al. (2015) as a common trait for many rivers in Central and Eastern Europe. The causes for these adjustments of fluvial styles consist in the climate changes and the subsequent dynamics of the vegetation cover. Moreover, shifts generated by tectonic movements overlap climate changes, thus leading to phenomena such as avulsion (e.g., the rivers of the Western Plain, Perşoiu et al., submitted), incision or floodplain aggradation.

As regards the LQ history of the Danube, data is still insufficient for clearly discerning between the role of climate changes (as relayed from the drainage basin through the tributaries) and that of eustatic oscillations at the confluence with BS. Several cores retrieved in the Danube floodplain downstream of

Giurgiu urban area (Fig. 4b, P and G points) have intercepted a massive layer of gravel and coarse sand dated to at least 32.7 and 15.9 ka (OSL dates) and seem to correspond to a braided river (Benecke et al., 2013; Hansen et al., 2017; Vespremeanu-Stroe, pers. com).

The depth at which the gravel layer was intercepted ranges between -17 m at Pietrele (Giurgiu), -14.5 at the junction with Mostiștea, -35 m at Hârșova, -25 m at Brăila, -12 m (just downstream of the junction with the Siret, but in the latter case the drilling stopped at the sands, Țuțianu et al., 2021), and -45 m at Tulcea and Sulina (the latter according to Giosan et al., 2009; Vespremeanu-Stroe et al., 2017).

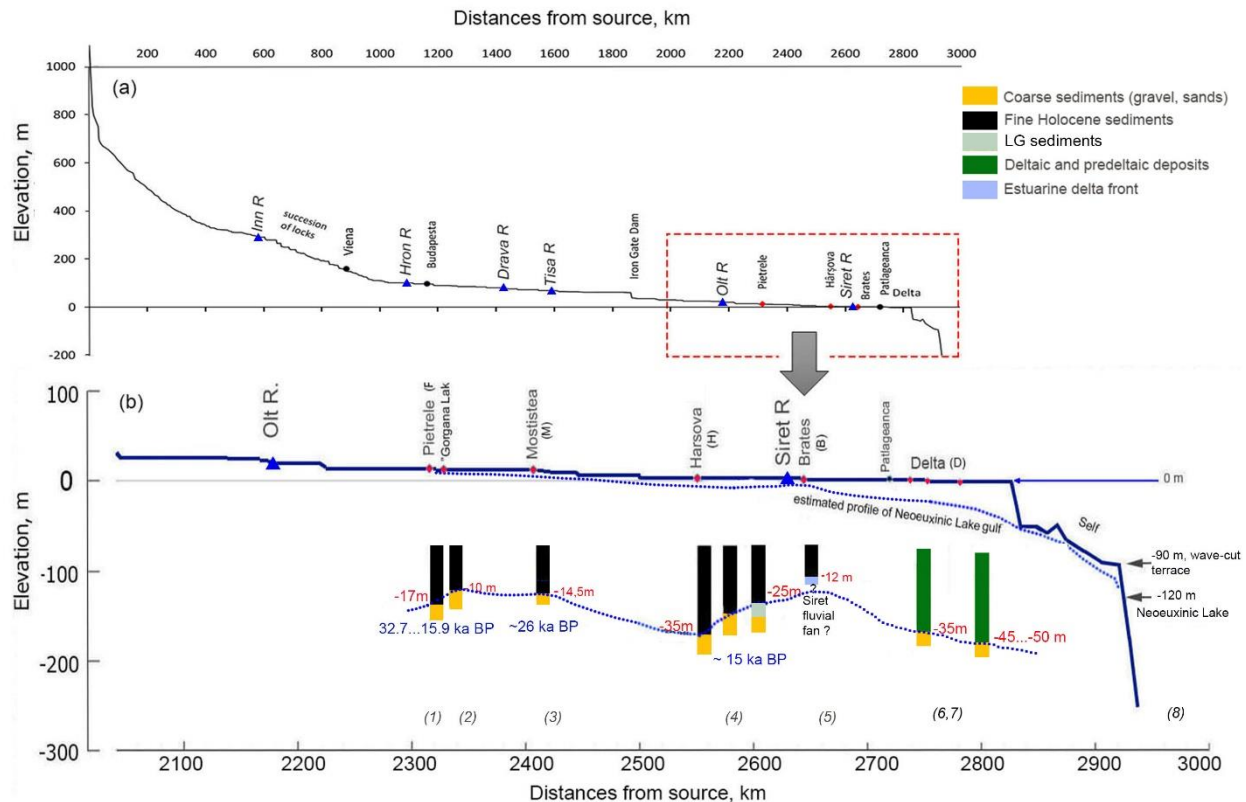


Figure 4 (a) Longitudinal profile of the Danube with marking of the lower sector; (b) Detailed longitudinal profile of the Lower Danube and approximate reconstruction of the neo-Euxinic Danube longitudinal profile (dotted line) with exaggerated drill representation along the vertical axis: (1) Benecke et al. (2013); (2) Hansen et al. (2018); (3) Vespremeanu-Stroe (personal communication) (4) Ghenea & Mihailescu (1991); (5) Tutuianu et al. (2018); (6,7) Vespremeanu-Stroe et al. (2017); Giosan et al. (2009); (8) Popescu et al. (2004, 2015)

The age and facies of the fluvial deposits indicate the course of the gravel-bed channel of the paleo-Danube during the neo-Euxinic phase, when the base

level of BS was more than 100 m below the present-day level (Fig. 4b). This course is irregular and appears as a succession of deeper and shallower reaches,

which is similar to the alternating riffle-pool morphology of current channels. One of the raised reaches of the coarse-grained bed of this course is located adjacent to the confluence with river Siret, thus suggesting a fluvial fan type of landform at the contact between the two rivers (although the authors of the Brateş drill estimate the depth of the coarse-grained bed at -25 m). We prefer to avoid speculating in this phase of the study, which focuses on collecting data; however, the mechanism through which the effects of base-level lowering are transferred upstream is thought-provoking (Posea, 2002 estimated that these effects were transmitted upstream as far as Giurgiu urban area), as well as the mechanisms of downstream transport (water + sediment) and the resulting morphology.

Relating these findings with data reported by studies carried out at Brăila Island (Ghenea and Mihăilescu, 1991) indicated that the lower floodplain continued to undergo an active process of aggradation with traces of avulsion, which advanced along with the anastomosed channel system of the BS inner shelf (Popescu et al., 2004; Lericolais et al., 2010; Popescu et al., 2015). The level rise of BS after 16 – 15 ka favoured the accumulation of finer-grained fluvio-lacustrine sediments corresponding to the deglaciation phase when the level of the Black Sea rose by ~40 m. This led to an increasing occurrence of the anastomosed typology of the river along with avulsion. As a result, the right fluvial branch which delimits Brăila Island, shifted to a

secondary status, whereas the left branch became the main one in the Early Holocene (Ghenea and Mihăilescu, 1991). This process was partially ascribed to tectonics due to the apparent overlap of the southern paleo-branch on the Peceneaga-Camena fault, a major crustal fault separating the north-Dobrogea orogen from the Moesian Platform (Popescu et al., 2004).

To the north of the Black Sea, the vast Eastern European Plain is crossed by several large rivers, among which the Dnieper stands out in terms of the amount of research data yielded by sediment archive-based reconstruction studies. This fluvial system was granted scientific attention due in no small part to the fact that Dnieper River was the only route linking the Black Sea to the decaying Scandinavian ice sheet. The occurrence of glacial meltwater pulses has been widely used to interpret the Black Sea hydrology during the Late Pleniglacial (Lericolais et al., 2011; Sidorchuk et al., 2011; Soulet et al., 2011).

The fluvial history of this drainage system was analysed by Borisova et al. (2006), but more in-depth by Panin et al. (2014, 2015, 2016), including the temporal succession of incision and avulsion processes throughout the past 25 ka (Fig. 5). The upper Dnieper valley formed after the ice cap retreat, which left behind specific residual moraine landforms (drumlins), subsequently shaped by fluvial activity during the Late Glacial and the Holocene.

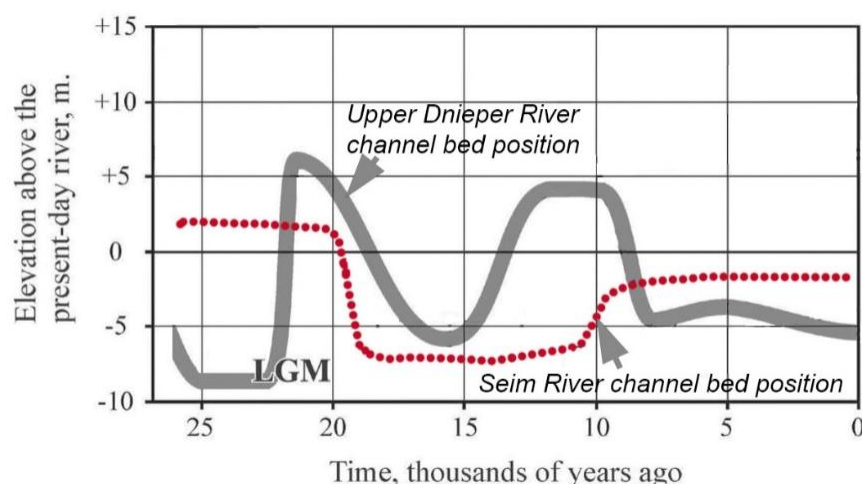


Figure 5 Incision/aggradation dynamics of the Upper Dnieper and Seim River since MIS 3, exhibited by changing position of the base fluvial deposits. (after Panin et al., 2014 and 2016 with additions)

The evolution of the river valley during and after MIS 2 was strongly influenced by proglacial effects, such as glacial damming and valley tilting due to forebulge development. This resulted in rapid aggradation around LGM, ensued by incision during Older Dryas (16.5 to 15 ka), and followed again by aggradation between 15.5 and 12 ka as a fluvial response to valley back-tilting continued in the Holocene. For comparison we included in Fig. 5 the channel dynamics of Seim river, a tributary of the Dnieper located farther to the south from the upper course of the Dnieper, in the middle sector of its drainage basin. This parallel leads to the conclusion that, within the same drainage system, channel aggradation and degradation are not necessarily synchronous for all rivers, and neither are channel planforms. In the case of the upper Dnieper, climate changes (expressed in terms of paleo-hydrology) have acted alongside glacial isostatic adjustment, resulting in valley aggradation and the evolution towards a braided style of channel during late MIS 2. This phenomenon was largely absent on river Seim, which lacks this type of control, and has thus developed a wandering sinuous style. The influence of glacial isostatic effects was probably responsible for the development of the East German rivers that demonstrated continuous aggradation since late MIS 5 until MIS 2 (Mol, 1997; Kasse et al., 2005).

Few paleo-geomorphological reconstructions based on sediment archives are available for rivers located south of BS. Studies focusing on the longest Anatolian river (i.e., Kizilirmak River), tributary of BS, have shown that its channel has undergone incision during the LGM, which was globally the coldest period of Upper Pleistocene. The timing of river response to climate changes was rather dissimilar in the case of Kizilirmak River compared to other European rivers, which have commonly incised their channels in warm-cold and cold-warm phase transitions according to the non-linear theory (Vandenberghe, 2008). This phenomenon was attributed to the relatively high streamflow discharge and the low sediment load, which resulted in augmented stream power, thus amplifying incision (Dogan, 2010). This reflects the contrasting climatic and environmental conditions specific to the two

regions (particularly the vegetation cover) both in the present and the past.

Additional paleo-fluvial information is available for River Sakarya (NW Turkey), whose history was deciphered based on the stratigraphy, geometry, and absolute dating of the four-step fluvial terrace staircases (Erturac et al., 2019; Erturac, 2021). Along a river reach located at ca. 50 km upstream of the river mouth, the authors have identified distinct erosional and depositional periods correlated with eustatic oscillations (Fig. 6). Between ~30 to 9 ka the T3 (+22 m) incision was documented as a consequence of a dramatic drop of BS base level (~140 m b.s.l), which forced the river to abandon floodplain formation and to deepen its channel by at least 25 m, cutting through the bedrock. Subsequently, the abrupt BS level rise recorded after 9.6 ka favoured the accumulation of sediment in the form of the future T2 terrace (+10 m), which would be incised later during the erosion phase at 1.8 – 1.2 ka. The evolution model established by the authors (Fig. 6) is based on several OSL dates, which indicate that the accumulation of terrace deposits occurred during major transgression phases (T3 of +22 m in Surozhian / MIS3; T4 of +55 m in Tarkhankutian / MIS5a), whereas incision and terrace carving took place in the Neoeuxinian / MIS2 and Post-Karangatian / MIS4-3. The latter processes were highly active in terms of their geomorphic effectiveness by achieving both the removal of previously deposited sediment material and the deepening of the channel through the bedrock. According to this model, incision was carried out during a stage when the Black Sea was isolated from the Mediterranean Sea and its level had dropped by 100-120 m. The limited geomorphological evidence and the scarcity of ages determined for the post-25ka period have prevented researchers from providing definitive answers to controversies regarding BS oscillations during this time frame.

To conclude, the comprehensive review of paleo-geomorphological studies based on fluvial sediment archives in the Black Sea drainage basin has led to the following general observations:

(i) the river response to major climate changes in the BS region was sensitive, although it lacks the high

resolution specific to other terrestrial records, such as speleothems or lacustrine sediments;

(ii) as the river size increases, along with the variety of water supply sources and their distribution across ample and diverse physical-geographical units, the sensitivity to climate changes diminishes (e.g., the Danube vs its Pannonian or Carpathian tributaries);

(iii) cold climates have been favourable for displacing and transporting large amounts of sediment from the source basins, such that river channels aggraded, assembled floodplains and evolved to braided channel patterns. The exception to this model was documented in the Anatolian Plateau, where the vegetation cover persisting during the cold phases prevented the massive transfer of

sediment to river channels, such that incision was a dominant event in cold climates;

(iv) by contrast, warm climates were conducive to vegetation growth, which contributed to a large degree to the retention of sediments displaced from the slopes, thus leading to meandering channels and, later on, to incision;

(v) regardless of the fluvial evolution model, other factors could disturb or subdue the role of climate changes, including tectonics (in the Pannonian Plain or the Western Plain of Romania), the vicinity to coarse-grained sediment sources (e.g., the medium-sized Carpathian tributaries), eustatic oscillations (as documented on the Danube or the Sakarya R.), or the accumulation of glacial sediment left behind by the retreat of the ice cap (e.g., on the upper course of the Dnieper).

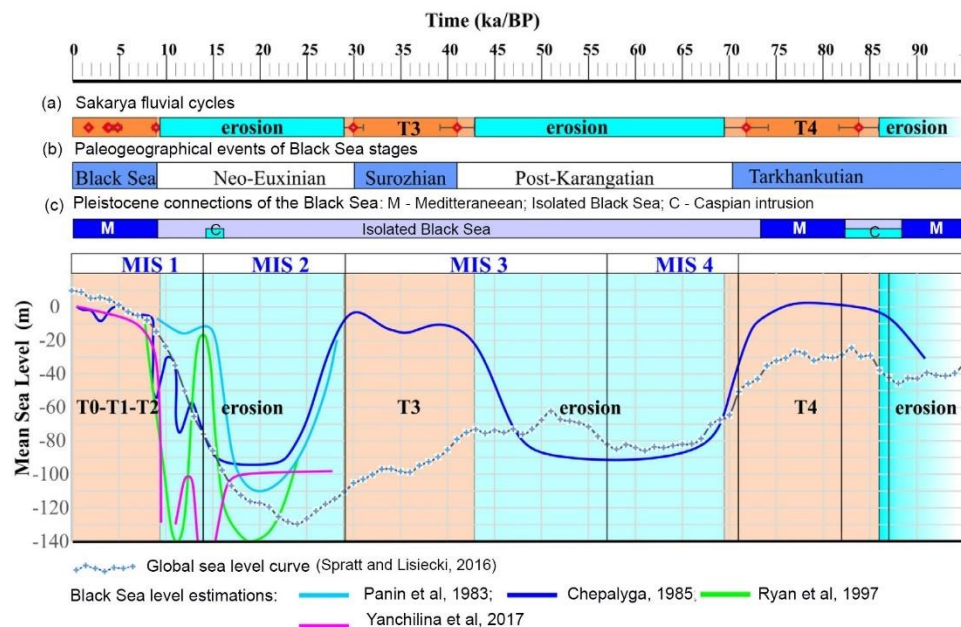


Figure 6 The correlation between the Sakarya fluvial cycles and the global and Black Sea level changes (after Erturac et al., 2019): (a) depositional and erosional periods with reported OSL ages and errors for Sakarya River during the Late Pleistocene T4 and T3, Holocene terraces T2, T1 and T0 (Erturac et al., 2019). (b) Main paleo-geographical events of the Black Sea according to Panin and Popescu (2007) and Yanina (2014). (c) Late Pleistocene connections of the Black Sea inferred from isotope studies (Badertscher et al., 2011)

4. Paleo-hydrological reconstructions based on paleo-channel morphology

Relict fluvial morphologies were used to carry out large-scale, highly accurate paleo-hydrological reconstructions in the Eastern European Plain, which is part of the BS drainage basin. The region accounts for 40% of the total area drained by the BS through

more than 72,000 streams, pertaining mostly to Dnieper and Don basins (to see Fig. 1). Within a latitudinal interval delimited by 57 N and 44 N, a significant number of relict paleo-channels were identified on the lower fluvial terraces or even in the floodplains (of which more than 700 river reaches were analysed, Sidorciuk et al., 2008). The majority of

paleo-channels fall into the meandering typology with amplitudes and wavelengths as high as 15 times larger compared to present-day meanders (Sidorciuk et al., 2001; 2009; 2011; Borisova et al., 2006). By applying a multi-disciplinary methodological

package (geomorphological, sedimentological, geochronological, palynological) on these relict morphologies, researchers were able to reconstruct the hydrological history of rivers in the region (Fig. 7).

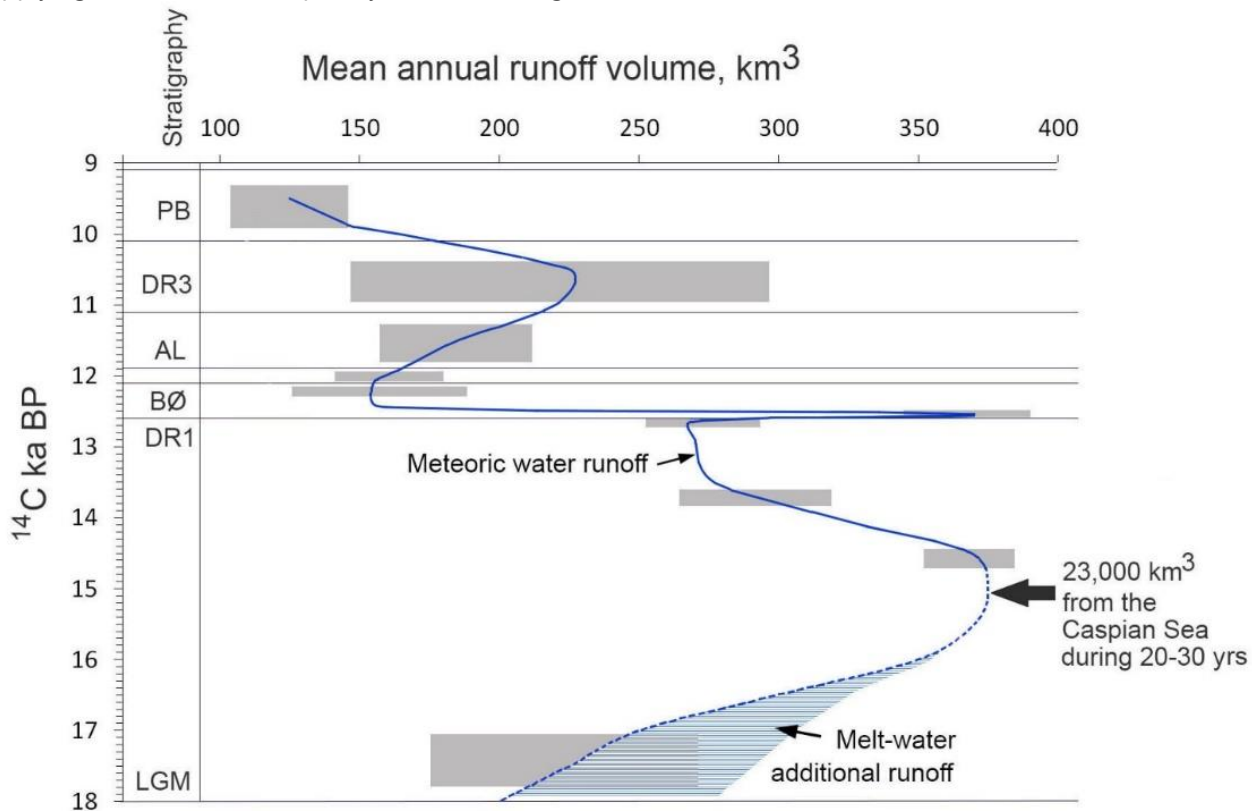


Figure 7 Reconstructed surface runoff in the Eastern European part of the Black Sea drainage basin since the end of Last Glacial Maximum (LGM) till Early Holocene. PB – Preboreal; DR3 – Younger Dryas; AL – Allerød; BØ – Bølling; DR1 – Oldest Dryas. Gray stripes-error of estimation (Sidorciuk et al., 2011)

The considerable differences observed between present-day channel morphology and past morphology point toward a significant change in the hydrological regime and the water discharge which occurred at the final stage of the last glaciation. The largest paleo-channels were active between 19 and 16 ka, i.e., shortly after LGM and towards the end of the last glacial (end of Pleniglacial) (Vandenberghe and Sidorciuk, 2020). The calculation method presented by the authors (Sidorciuk et al., 2011 and references therein) allowed for the spatial reconstruction of surface runoff in the basins of Dnieper and Don and the estimation of the streamflow discharge received by BS from these tributary basins. The peak runoff value recorded during the pre-Oldest Dryas era was documented in

the area adjacent to the ice cap (ca. 600 mm) and decreased southward to ca. 200 m. The maximum amount of discharged water was up to four times higher compared to present-day values, with rivers from the Eastern European part of BS basin equating a total amount of 365 km³/year. Subsequent to this hydrological peak recorded around the end of the Pleniglacial, streamflow discharge decreased overall, albeit at higher values compared to current discharges. The present-day discharge values of rivers pertaining to BS drainage basin were reached at the beginning of the Holocene.

A similar method was employed to reconstruct the volume of water overflow from the Caspian Sea to the Black Sea through the Manych channel. During the hydrological peak, Volga River discharged into

the Caspian Sea $\sim 500 \text{ km}^3/\text{year}$, leading to the Khvalynian transgression, which, for a short period of time, resulted in the communication between the two seas and the subsequent transfer of up to $23,000 \text{ km}^3$ of water into the Black Sea (Fig. 7).

Hydrological events from the Eastern European part of the BS basin are described chronologically as follows (Fig. 7): (a) 18-15 ka was a cold period with lower precipitation compared to the present, albeit the runoff coefficient was higher due to reduced evapotranspiration. River discharge amounted to $200\text{--}300 \text{ km}^3/\text{year}$ in total, including an additional input of meltwater (up to $100 \text{ km}^3/\text{year}$) from Fennoscandinavian ice cap decay and proglacial lakes; (b) 15-14 ka was a warmer, more humid interval, with higher precipitation compared to current values and peak runoff values ($300\text{--}365 \text{ km}^3/\text{year}$), thus resulting in the formation of the largest meandering rivers. During a short interval (of a few decades), a large amount of water was discharged from the Caspian Sea into the BS; (c) 14-10 ka was a period marked by low magnitude climate oscillations which corresponded to runoff decrease. At the end of this interval, large paleo-channel meanders transformed into smaller meanders. In the Holocene the runoff diminished to $100\text{--}110 \text{ km}^3/\text{year}$, which continue to be specific until present.

Similar assessments on the hydrological history of the Danube have not been carried out to date. The drainage basin of the Danube accounts for 2/3 of the total BS basin and discharges into the BS a water amount of ca. $200 \text{ km}^3/\text{year}$, which is twice the water input discharged by the Eastern European tributaries. For example, Sidorciuk et al. (2011) estimated that during the LGM and LG, Carpathian tributaries had mean water discharges up to three times larger compared to the present and similar variations to rivers from the Eastern European part of the BS basin, with its maximum shifted toward the beginning of the LG.

As a final observation regarding paleo-hydrological reconstructions, it should be noted that the high surface runoff was specific throughout the entire LGM – LG interval, at least in the Eastern European part of the BS basin (of more than $350 \text{ km}^3/\text{year}$) compared to current values (of $110 \text{ km}^3/\text{year}$, reached during the early Holocene). The

available hydrological reconstructions reviewed in this study provide no support for the “flood hypothesis” which asserts that BS sea water level was $\sim 120 \text{ m}$ lower prior to the connection with the Mediterranean Sea (at ca. 9.4 ka). Novel data regarding the formation of the subaerial Danube Delta is strong evidence of the sea water level position at ca. 30 m below the present BS level prior to the reconnection with the Planetary Ocean (Giosan et al., 2009; Vespremeanu-Stroe et al., 2017). The level oscillations of the Black Sea and their reconstructions are approached in detail in the following section of this study.

5. Reconstructions of Black Sea level oscillations and effects on paleo-drainage on the continental shelf

The Black Sea is a massive, deep-water body that communicates with the World Ocean through the Bosphorus Strait, whose uppermost threshold is at 35 m depth. The international expeditions from the 1960s and 70s and the seminal work of Degens and Ross (1974) were among the first results on the Late Quaternary history of BS that fascinated the geoscientist community and triggered further interest for knowledge which persists to this day. In addition to bottom waters anoxia and low salinity, another captivating revelation was related to the oscillations of BS water level and their implications. The last 25 ka stand out as the best documented time frame in this regard. During the Upper Würm glaciation (Late Weichselian, Late Valdai), which corresponds to the Neoeuxinic BS phase, the level dropped dramatically at depths ranging between $\sim 90 \text{ m}$ (Chepalyga, 1984) and $\sim 140 \text{ m}$ (Ryan et al., 1997).

Hypotheses diverge not only on the maximum depth reached during BS water level drop but also on the alternation of oscillations during the 25-ka time frame. Fig. 8 sums up some of the most discussed hypotheses (ranked by Yanchilina et al., 2017 into four categories, as previously shown in the introduction), reflecting the variety of interpretations based on the data collected to date. Moreover, we added the runoff volume curve for rivers from the Eastern European part of BS basin (which amounts to ca. half of the volume of water discharged by the Danube) as reconstructed by Sidorciuk et al. (2011)

to gain a better understanding of the link between BS level oscillations and runoff in the drainage basin. However, this study does not aim to examine these disputes in-depth but is instead focused on discerning the role of BS as a base level for fluvial processes of the tributary rivers based on the scarce data available to date.

Regardless of the supported hypothesis, it is certain that the coastline has retreated far from its current position, particularly in the NW, where a large part of the continental shelf was exposed. The BS shelf accounts for 30% of the sea surface area (Panin and Jipa, 1998) and its maximum extension (over 190 km wide) lies in the NW part, between Crimea and the Danube Delta. In front of the Romanian coast the shelf is 50 to 100 km wide. This large area was exposed during the Neoeuxinic period, therefore tributary rivers deepened into the newly exposed terrain to ca. 90 m on the outer shelf. Buried channels were observed since the earliest investigations at Kerch strait and Azov Sea (where a 5-km wide, 62-m

deep valley with gravel bed channel in “a single alluvial complex consisting of the channel, oxbow and floodplain facies” attributed to the paleo-Don was documented - Skiba et al., 1976, cited by Ryan et al., 1997). Furthermore, several international and national expeditions identified on the Ukrainian and Romanian shelves similar submerged relict morphologies buried by recent sediments.

Figure 9 shows the map of reconstructed paleo-geomorphologies within the Romanian shelf (Fig. 9a) based on the data provided by several Romanian-French and Romanian expeditions from 1998 to the present (Popescu et al., 2004; 2015). The continental shelf is composed of an inner shelf delimited by the 27 m to 30 m isobaths and slope gradients ranging between 1.1° and 4°, continued by an outer shelf with slope gradients below 1°. On both shelf units were identified numerous channel beds with widths ranging from 400 to 1500 m and varying depths of 20 to 90 m, buried under a thin layer of fine sediment moulded on their contours.

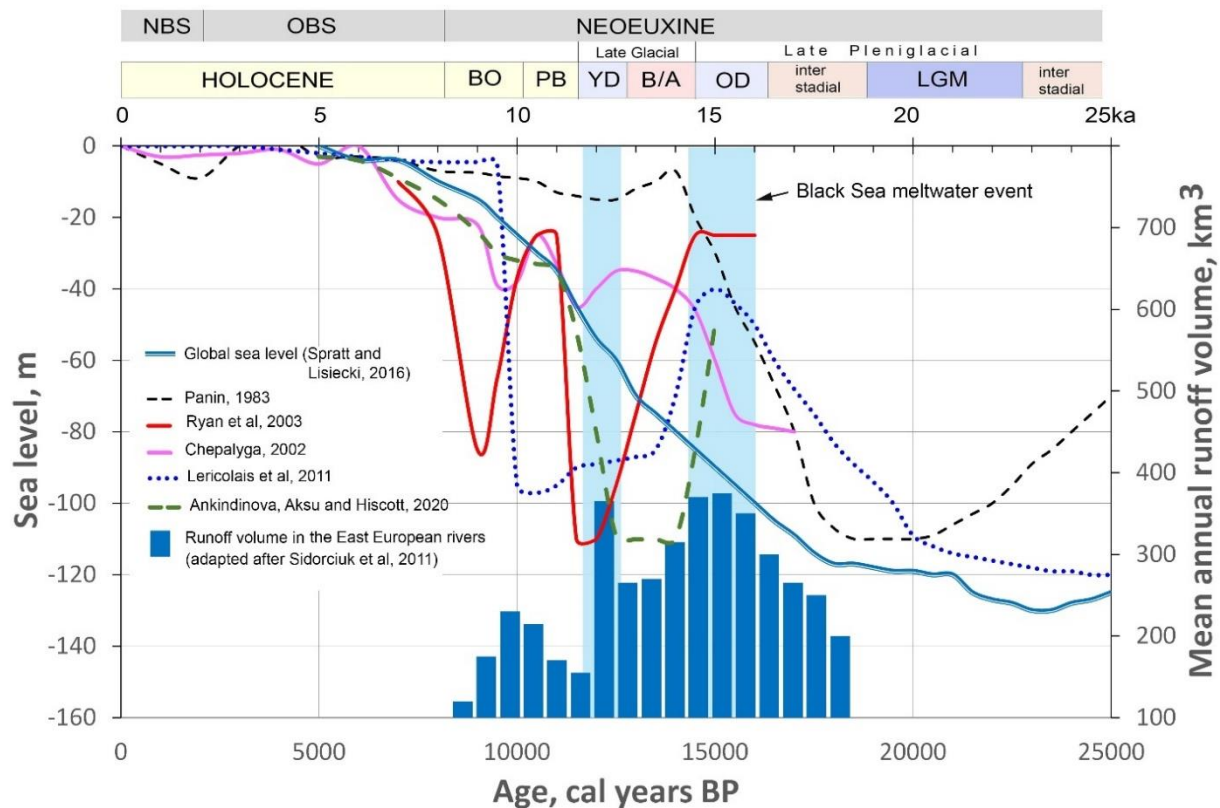


Figure 9 Sea water level reconstructions for the post - Glacial to Holocene Black Sea: NBS - New Black Sea; OBS - Old Black Sea; BO - Boreal; PB - Preboreal; YD -Younger Dryas; BA - Bolling/Allerod; OD - Oldest Dryas; LGM - Late Glacial Maximum. Reconstructed runoff in the Eastern European part of the Black Sea drainage basin is shown after Sidorciuk et al. (2011)

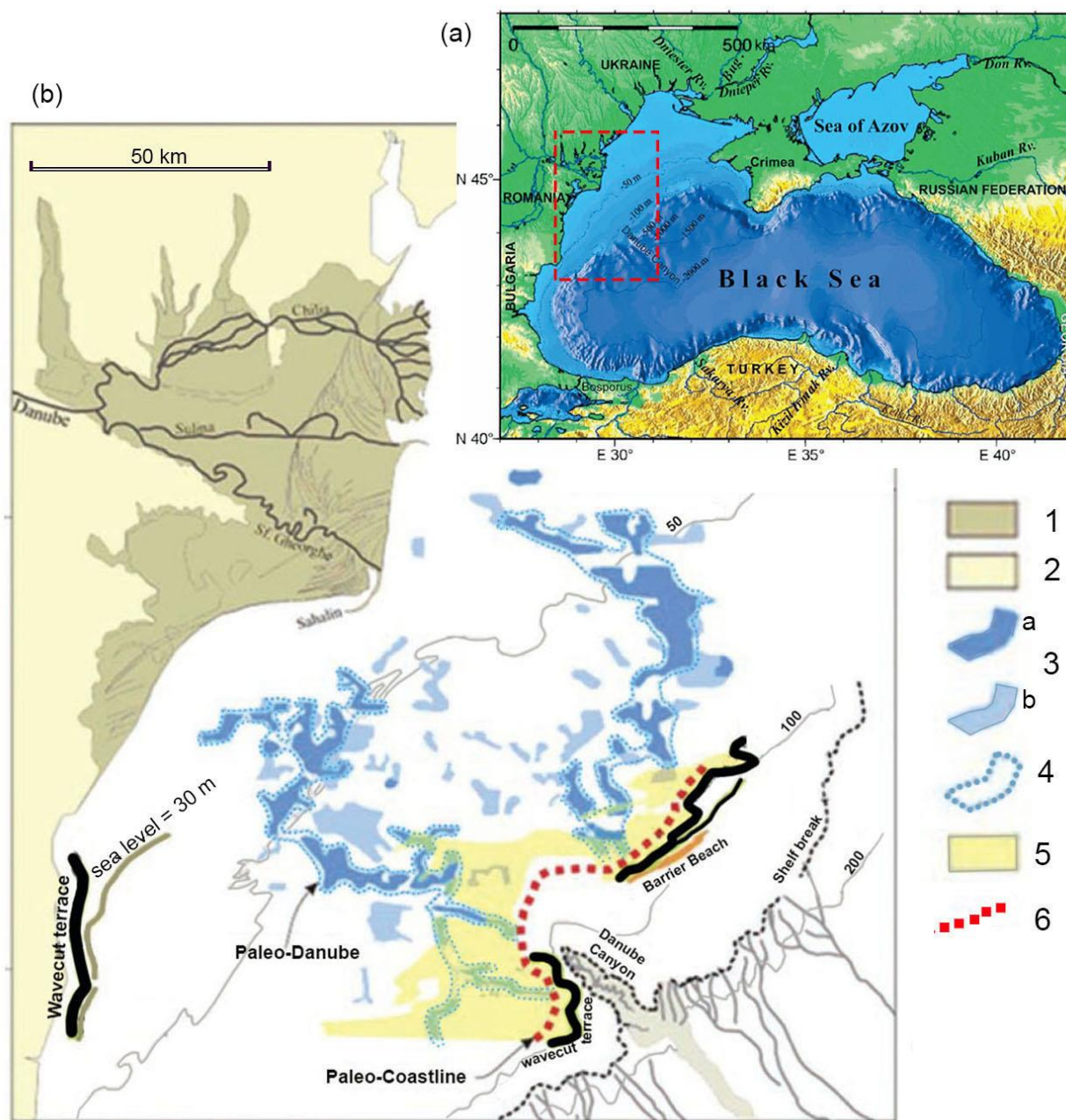


Figure 9 Schematic geomorphological maps of the Black Sea shelf (redrawn after Popescu et al., 2015 and Caraivan et al., 2017): A. Location of the mapped area; B. Romanian shelf: 1, Danube Delta (Holocene); 2, Pre-Holocene deposits; 3, Areas of the buried river channel beds with different degrees of density (a. higher; b, lower); 4, Probable outline of two anastomosed river channel systems; 5, Dunes; 6, Paleo-Coastline

Their stratigraphic position lying directly under the discontinuity at the base of the Holocene suggests they formed during the last lowstand. The distribution of buried channels shows clusters oriented in two directions which indicate distinct drainage systems (Fig. 9b). The sinuous-anastomosed pattern could point to the paleo-

Danube course, which divided into several branch, similar to the present-day Danube Delta. The southern drainage system continues with the Danube canyon, whereas the origin of the northern one is uncertain. This could be a second branch of the Danube. The two rivers (or branches) end their course at the paleo-coastline, which was probably the base

level during the Neoeuxinic period. The paleo-coastline was the contour of an old gulf where the two river branches discharged their waters through a canyon named "al Dunării" or Viteaz. This canyon was entirely submerged and located below the wave action zone, as attested by the position of wave-cut terrace (Fig. 9b). According to the authors (Popescu et al., 2004) it formed due to the low BS base level and the significant sediment input via the paleo-Danube.

To conclude, the circum-Pontic fluvial activity was controlled in part by the BS level during some time frames. For the Neoeuxinic period are notable the reconstructions of sinuous-anastomosed channels of major rivers (Danube, Dniester, Dnieper), as well as for smaller tributaries deepened into the extensive surface of the north-western BS continental shelf. The Danube Canyon incised by the river under submerged conditions, as it appears, along a regional fault line (Pecineaga-Camena) provides a spectacular phenomenon. However, we were unable to find evidence regarding the upstream distances to which the effects of eustatism were encountered along river valleys. Therefore, the discussion remains in the hypotheses stage, such as the knickpoint migration theory. Armas et al. (2018) identified a succession of four major knickpoints along Argeş and Dâmboviţa rivers (tributaries of the Danube ca. 500 km upstream of the BS confluence) which can be correlated with terrace profiles, thus proving "that the influence of marine regressions and transgressions during glacial and interglacial stages, respectively, was apparent active up to Bucharest" (page 15). This hypothesis requires more solid evidence in its support, especially considering that other investigations of present-day fluvial activity concluded that upstream effects of base level oscillations are rather limited (Leopold and Bull, 1979). Moreover, lab-based and theoretical models show that the impact of base level drop in terms of river incision is maximal in its early stages in the confluence area and decreases markedly upstream (Begin et al., 1981; Begin, 1988).

6. Final remarks

Rivers have a heritage, according to Luna Leopold (1915 - 2006), a classic of fluvial geomorphology, but we know nothing of their beginnings. Simply put,

"there is a continual modification, but at any stage the existing system is different from that which existed at a previous state, but has been influenced by that earlier stage" (Leopold and Bull, 1979, page 169). Some authors compare this overlay to a palimpsest, with the sole exception that more recent "writings" are influenced by previous ones, thus providing evidence of evolution. While we are aware that the legacy of rivers is not highly accurate in terms of recording the changes of external stimuli, the sensitivity of their response is well documented. In this case, our contribution consisted in assembling information regarding the legacy of circum-Pontic rivers (Fig. 10) where the focus is placed on the dynamics of fluvial styles, the most visible traits of rivers. The vertical axis indicates the time/age (between 25 ka and 8.5 ka), while the horizontal axis shows circum-Pontic rivers depicted as their channel planforms. On each side of the river diagram are shown the temporal variabilities of two major controls which exerted significant influence during the analysed time frame, namely the climate (reconstructed based on ice core records and other terrestrial records from BS basin) and BS base level oscillations. Thus, we postulate that rivers were constrained from two diametrically opposed directions to which they were forced to respond: to the control exerted by the drainage basin originating upstream and to the downstream control of the oscillating base level.

We illustrated the climate variability at the drainage basin scale using the high resolution, accurately dated NGRIP diagram (Greenland ice records; Rasmussen et al., 2014), which was integral to establishing the timeline of climatic phases in the Northern Hemisphere (Moreno et al., 2014).

From the geographical area of the BS basin, we used the time series of two terrestrial records: the oxygen-isotope series of two stalagmites from Sofular Cave, in NW Turkey, which is regarded as the record with the highest temporal resolution both for the Holocene and the pre-Holocene (Kern et al., 2019; Burstyn et al., 2019), and the reconstruction of mean annual temperature and precipitation based on pollen analyses of a sediment and peat sequence retrieved at lake Preluca Tiganului, Gutâi Mts, in NW Romania (Feurdean et al., 2014). Both terrestrial

records used for this study fit reasonably well with the NGRIP series.

The rivers analysed in this study were grouped into four morpho-structural units which, in turn, illustrate the nuances of river channel responses to the controls mentioned previously. The first group comprises the rivers of the Pannonian Basin, among which river Tisa stands out in terms of size and has been the object of numerous paleo-evolution investigations. The second group includes the rivers draining the Romanian Carpathians. Both of these clusters are tributary to the Danube. The third group consists of the rivers draining the Eastern European / Russian Plain, whereas the fourth one includes just two southern BS tributaries from the Anatolian

Plateau and the Pontian Mts. As previously stated, we have taken into consideration only the rivers for which paleo-evolution data was available and can be characterized as circum-Pontic to some degree.

The non-linear model of river cyclicity as a response to climate changes shows that relatively short phases of fluvial instability occur mainly at the transition from warm (temperate) to cold (periglacial), and from relatively cold to warm periods (Vandenberghe, 2008). Instability is manifested early on through vertical fluvial processes (e.g., a succession of incision or vertical erosion and sediment accumulation). Vertical instability is ensued by channel planform changes and floodplain development during the long stable phases of cold or warm periods.

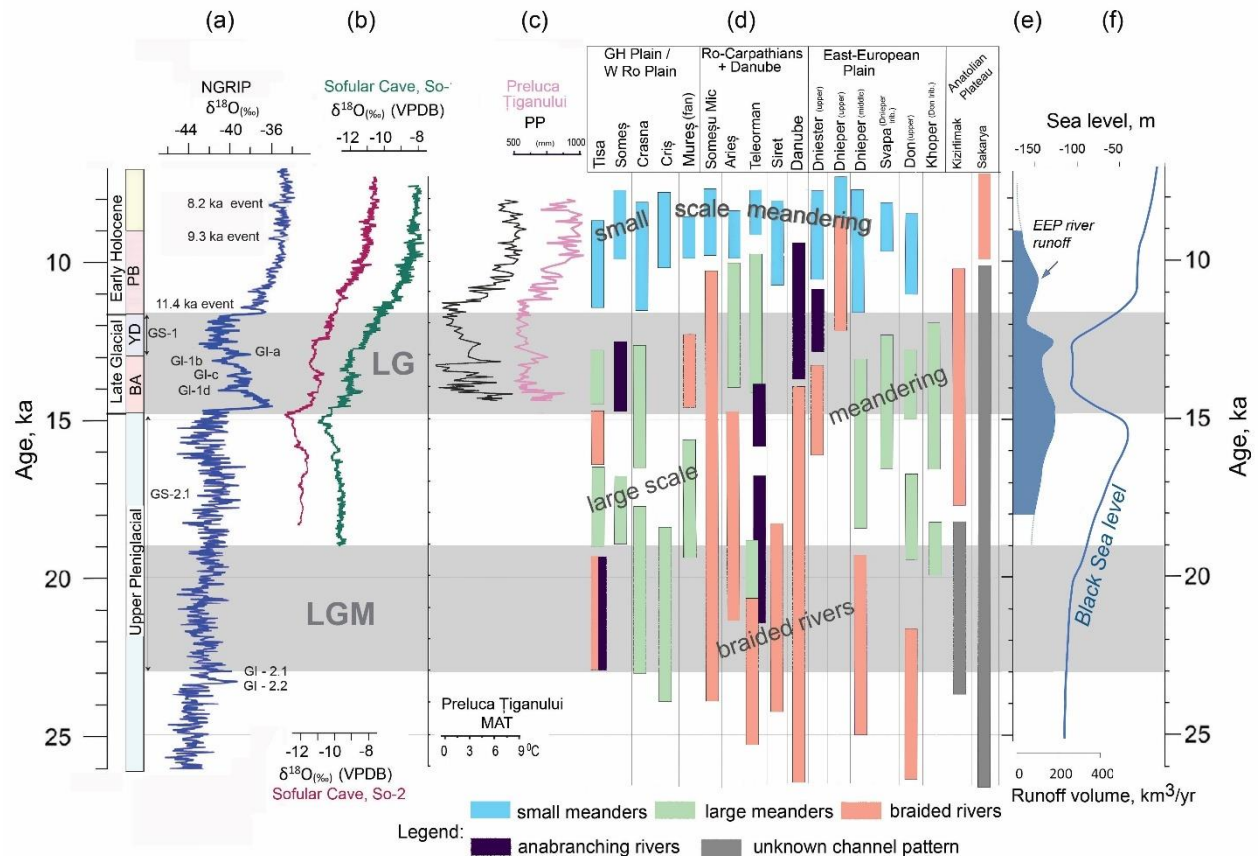


Figure 10 A simplified model of fluvial pattern evolution for circum-Pontic rivers in the Late Quaternary and correlation with paleoclimatic records, river flow dynamics and Black Sea level oscillations: (a) NGRIP $\delta^{18}\text{O}$ curve; GI – Greenland Interstadial; GS – Greenland Stadial (Rasmussen et al., 2006); (b) The $\delta^{18}\text{O}$ time series of stalagmites So-1 and So-2 from Sofular Cave, NW Turkey (Fleitman et al., 2009); (c) A pollen-based quantitative reconstruction of paleoclimate from NW Romania (Feurdean et al., 2008); (d) Circum-Pontic rivers grouped by morpho-structural units of the Black Sea drainage basin; (e) Reconstructed surface runoff in the Eastern European part of the Black Sea drainage basin (Sidorciuk et al., 2011); (f) Water-level fluctuation in the Black Sea since the Last Glacial Maximum (Ankindinova et al., 2020)

This model, which is widely recognized in geomorphological reconstructions, is fundamental for our approach regarding the evolution of the fluvial system as presented in this study: elongated stripes indicate stable phases, interrupted by interphase changes.

The following section sums up the dominant traits of fluvial styles during the climate phases of the 25 – 8.5 ka period and the time frames when the changes were documented (Fig. 10).

(i) The Upper Pleniglacial (28 - 14.65 ka cal BP), or, more specifically, the Late Glacial Maximum, LGM (23 – 19 ka), was a cooler, drier period, marked by the largest southward extension of permafrost, the broadest advancement of mountain glaciers and Scandinavian ice cap, and a significant accumulation of loess. River channels were forced to manage oversized sediment yields which resulted in increased river gradients and the formation of braided river channels, to strong wind activity and dune formation within floodplains were added (Kasse et al., 2010).

Braided channels were the dominant type of fluvial style. From this period, a continuous sheet of gravel and coarse sand is preserved in the structure of low floodplains, mainly in Carpathian rivers, including the Danube, but was also reported in Pannonian or Russian Plain rivers. By contrast, rivers from the Anatolian Plateau showed opposite tendencies by deepening their single channels, which are indeterminate in terms of typology (Dogan, 2010). At the contact with the BS base level, circum-Pontic rivers were bound to carve their channels to the confluence with the sea whose level was, at the time, 100 to 120 m lower compared to present. As such, vertical erosion was the prevailing process, with rivers cutting through their beds as deep as the bedrock in some cases, as was documented on the Sakarya R at 50 km upstream of the BS level. Around the same time, in mid Neoeuxinic phase, the paleo-Danube downstream of Brăila had deepened by 30-35 compared to present, as indicated by drills which intercepted the gravel sheet deposited during the Neoeuxinic phase. It is conceivable that the effect of the base level drop has been propagated upstream as far as Giurgiu urban area (see Fig. 4b). As regards the horizontal plane, sinuous-anastomosed channels formed particularly in the NW plain of the BS

(currently acting as continental shelf), as well as valleys more than 60 m (likely the paleo-Don canyon) to 90 m deep on the outer shelf (paleo-Danube canyon. In other words, the discussion on the upstream distance to which the effects of the base level drop are propagated remains open, particularly for major rivers such as the Danube.

(ii) Towards the late part of the Upper Pleniglacial (16 – 15 ka), due to the warmer, more humid climate conditions, the surface runoff reached the highest recorded values, leading to an increase in mean annual discharge of rivers by 3 to 4 times. The highest increase was documented for the maximum discharge, recorded in spring (i.e., 6 to 7 times higher than current maximum discharge values), and these were instrumental for shaping macro-meanders, as ample as 15 times larger compared to present meanders on rivers of the Eastern European Plain (Vandenberghe and Sidorciuk, 2020). Fluvial systems were subjected to massive transformation, evolving from braiding to large-scale meandering, which was mainly documented in rivers from the Pannonian Basin and the Eastern European Plain. As regards small and medium-sized Carpathian rivers, the braided style persisted throughout this phase, possibly due to the proximity of coarse-grained sediment sources and the preservation of a high energy environment (e.g., the well-documented cases of Someșul Mic river channel in the Transylvanian Depression – I. Perșoiu, 2010 or Moldova and the middle course of river Siret, all under the strong control of Carpathian sediment sources - Ichim and Rădoane, 1990; Rădoane et al., 2008; Chiriloaei et al., 2012; Rădoane et al., 2015). In the case of rivers influenced by the retreat of the ice cap a lack of synchronization of river processes (aggradation-degradation) has occurred, with effect on changes in river styles. The examples of the Upper Dnieper and the downstream tributaries are suggestive in this respect (Panin et al., 2014, 2015, 2016).

During this period, the BS receives the largest amount of water through its tributaries as well as from the Caspian Sea. The sea level rises from -120 m to -40 m, but the BS is still not linked to the Mediterranean Sea (see Fig. 8). All the confluences with the sea are flooded and sedimentation became

the dominant process due to the sea level rise. The model elaborated by Erturac et al. (2019) for Sakarya river (see Fig. 6) shows that phases of sea level rise were favourable for the accumulation of fluvial terraces at ca. 50 km upstream on tributary rivers. Some reviews (e.g., Blum and Tornqvist, 2000), as well as certain empirical studies have indicated that the distance to which the effect of base level rise is propagated upstream is limited to less than 10-15% of the river length (Leopold and Bull, 1979), with factors such as river gradient and composition of bed material accounting for the variation (N. Rădoane, 2002).

(iii) The Late Glacial, LG (14.65 - 11.65 ka cal BP) was characterized by gradual warming with fluctuations in the dynamics of the forest cover which expanded from the east, and by diminishing permafrost. The rapid cooling of the Younger Dryas caused a fragmentation of the boreal forest cover and a re-expansion of steppe-tundra habitats, as well as drop of the upper treeline from 1000 to 700 – 600 m a.s.l. in the Carpathians (Feurdean and Tanțău, 2017). It was ensued by the Early Holocene (11.65 - 8.5 ka), characterised by fast warming which reflected in the expansion of temperate deciduous forest. The channel response consisted in adapting to the streamflow discharge decrease (from 350 km³/yr to less than 250 km³/yr for EEP rivers, see Fig. 10d, and similar trends for the entire BS drainage basin). The channel planforms were constrained as large-scale paleo-meanders transformed to smaller meanders, while both the channel depth and width decreased. Simultaneously, gradual incision of the floodplains occurred in the context of meandering systems. This stands out as the second largest transformation undergone by rivers during their LQ history, after the major shift of LGM and LG, and was mostly ubiquitous throughout the entire BS basin, as illustrated in Fig. 10. However, a contrary tendency was documented in the Anatolian basin of river Kızılırmak. The warming climate of the late LGM generated low precipitation and scarce vegetation which led to channel bed aggradation (with sediment thickness reaching 18 m along the main valley) and the evolution to a braided channel (Dogan, 2010). As regards the BS level oscillation, we have mentioned previously that this period has been heatedly

debated and hypothesised on. We have opted to illustrate in Fig. 10e the curve estimated by Ankindinova et al. (2020) since this is the most recently published and is based on the latest field data. The river response to sea-level oscillations during and post-LG has not been determined accurately in case study investigations, therefore this topic remains open to further research. The best-documented case remains Sakarya river (Erturac et al., 2019) which has been discussed in section 3.

(iv) The Early Holocene was marked by a rapid BS level rise (at a rate of 10 m/100 yrs, such that in 1000 years, the BS reached the same level as the Marmara Sea, cf. Lericolais et al., 2017). The sea water flooded large areas to the NW of the BS and advanced upstream on the tributary valleys. The effects of the base level rise were reflected in the evolution towards the anastomosed channel style, for which the Danube is the best example.

To conclude, this is the story that could be pieced together based on the data regarding the legacy of circum-Pontic rivers available at this time. Aside from the general trend consisting in the dominant control exerted by climate, river channels were exposed to additional influences, including tectonics, the effects of certain types of vegetation in relation to the control of runoff and discharge, and most importantly on sediment displacement, the obstruction by glacial sediments left behind by the ice cap retreat, the river size and proximity to coarse-grained sediment sources, all of which were briefly analysed in documented case studies. As regards the role of Black Sea level oscillations and the response of tributary rivers, this topic requires further in-depth paleo-evolution investigations, which could shed light on the unknown facts which are fuelling the debate. To date, little information has been published on the fluvial history at the contact with the sea, with the exception of the Danube Delta and Lower Danube valley to some extent, but even in this case, more data is necessary to decipher the events of the Early Holocene, whether the connection with the Mediterranean Sea was abrupt or slow and what kind of sedimentary imprints were left by tributary rivers in this time frame (Giosan et al., 2006, 2009; Vespremeanu-Stroe et al., 2017; Țuțuianu et al., 2021). Hiscott and his team, who are known as

authors and major supporters of the Outflow Hypothesis, concluded an important chapter of their work published in 2007 by stating that “more work is

needed to understand fully the evolution of the (...) Black Sea”, which is widely applicable in research, but more so in the case of such an unusual, intriguing sea.

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Annex 1 Compiled data on the circum-Pontic river response to climate and sea level changes during Late Quaternary

	River	Before 14.65 ka Upper Pleniglacial	14.65 - 11.65 ka Late Glacial	11.65 - 9.4 ka Early Holocene	9.4 - 8.5/8 ka Early Holocene	Source
1	Siret, Romania, (dw. Suceava cf.)	20.3ka, aggrading, braided river, coarse gravel	Large scale meanders, estimated water discharge = 5 – 6 times the current discharges	Gradual incision, meandering	9.4 ka medium gravel lobes of lateral accretion	Rădoane et al. (2015, 2018a, b)
2	Moldova (up. Siret cf)			Braided, boulder and coarse gravel at the base of T 4, woods 10...9.5 ka		Chiriloaei et al. (2012); Rădoane et al. (2018b)
3	Someș (in D. Transilvania)			Braided, coarse gravel at the base of T 4 m, woods 9.8ka A generalized channel incision; single channel with large meanders by the maintenance of a still significant discharge; avulsion to NW		Rădoane et al. (2018a)
4	Someș (in Western Plain, up Tisa cf)	28.3 – 15.2 ka – large scale meanders, channel incision	A multi-branching channel system oriented to NNW			I. Perșoiu et al. (2022, submitted)
4a	Crasna (in Western Plain)		Large scale meanders, 13.6 ka avulsion to the NW		Small scale meanders	
5	Someșu Mic, Someșu Cald (in D. Transilvania)	28 ka, meandering pattern, incision of the T8-15 m 25-24.3 ka, anastomosed channel	24.3...11.4ka coarse gravel-bed braided channel, aggrading	10.6...9.7 ka channel metamorphosis from a braided into a meandering pattern; underfit river	Small scale meanders	Perșoiu (2010)
6	Arieș (D. Transylvania)	37.8 ka, T20 alluviation 18.8ka coarse gravel, braided river	14.6-12.9 ka, meandering pattern			Perșoiu and Perșoiu, in prep.
7	Mures (alluvial fan, up Tisa cf)		Braided, frequent avulsions, 3 – 4 times higher	a multibranched system on the alluvial fan	Meandering, 1.5-2.5 times the current discharges	Kiss et al. (2014)

8	Crişuri (W Plain, GHPlain)	Meandering pattern	discharge than nowadays A multi-channel braided fluvial system, neotectonic uplift			Nador et al. (2011)
9	Tisa (NE of GHPlain)	40-33 ka, braided; 33-32ka -large meanders	Large meanders: 5 – 8 times higher discharge than nowadays	Large meanders; channel vertical incisions and avulsions	Decrease of large meanders	Gabris and Nador (2007); Gabris et al. (2012); Kasse et al. (2010)
10	Danube (Iron Gates)			9.5-9 ka settlement abandonment is linked to increased flooding along the Danube	8.2-7.9ka, settlement abandonment is linked to increased flooding along the Danube	Bonsall et al. (2002; 2014)
11	Teleorman (tributary of the Danube, Danube Plain)	37 ka BP coarse gravel braided channel 21.6ka anabranching channel	12.8ka Large scale meanders (up to five times larger than the present ones)			Howard et al. (2004); Macklin et al. (2010)
12	Danube (lower course, Giurgiu-Oltenița)	32.7 ka....15.9ka layer of sands and gravels interpreted as corresponding to a braided river		9.3 ka Limnic phase, interpreted as being linked to the rapid increase in the Black Sea level in the early Holocene		Benecke et al. (2013) Nowacki et al. (2018)
13	Danube (lower course, Mostiște valley)	25ka...20ka layer of sandy gravels with a sandy layer on the top, probably corresponding to a braided river	18ka...8ka Intercalations of thin layers of sands, silt and clay, probably corresponding to a anastomosed fluvial environment	7.5 ka Limnic phase		Vespremeanu-Stroe (personal communication)
14	Danube (Dunărea locality)			8ka Intercalations of fine sediments		Vespremeanu-Stroe (personal communication)
15	Danube (Brăila island) No numerical ages	During this period, the northern part of Braila Island functioned as a gulf of the Black		11 ...9.5ka probably corresponds with a new coarse alluvial sequence reported in the		Ghenea and Mihailescu (1991)

		Sea, extending to the town of Braila		northern part of Braila Island, suggesting the reinstallation of the Danube as a river in this area.	
16	Brateş Lake, Danube-Prut rivers			>8ka, estuarine delta front 8-7.9ka, foreshore 7.9 – 5ka, river floodplain <5 ka, lake	Țuțianu et al. (2018)
17	Upper Dnister, Ukraine	25-18ka, braided channel, T 5-10m, 20m sediments in terrace	13,8-11 ka, traces of oldest meander with widest radius, anastomosing/meandering	8.5-7.8 ka, meandering, first distinct paleomeander	Huhmann et al. (2004)
18	Upper Dnister, Ukraine	15 ka, braided channel, T 5-6 m			Gebica et al. (2016)
19	River Dnieper, upper reach	Significant aggradation occurred as a fluvial response to valley back tilting caused by forebulge collapse	large palaeochannels were active in the Early Holocene between 12 and 8.5 ka; They were formed by flood discharges ca. three times larger, and with a valley gradient some 30 % higher, than those of the present-day river	deep river incision to the modern levels occurred between 9 and 8 ka; the channel type changed, from large braided or single-thread to sinuous	A.Panin et al. (2014) A Panin et al. (2015a, b)
20	River Seim, middle Dnieper, Ukraine - new datings	Braided channel, high level flood Incision 18-14ka	high level flood, large meanders	10 – 9 ka, small meanders	A Panin et al. (2016)
21	Seim River, Svapa River, middle Dnieper, Ukraine	16-14ka width and meander wavelength are 7-10 times greater than those of the modern channels 17 ka ago were abandoned	formation of smaller channels, climate warming	relative stability of hydrological regime and small variations of water runoff	Borisova et al. (2006)
22	Dnister, Dniپر, Don, Bug –	18-15ka, cold and dry, precipitation	15-14 ka, warmer and humid (surface runoff		Sidorchuck (2003)

	Russian Plain rivers	lower than today (surface runoff = 210 km ³ /yr)	=300-365 km ³ /yr), large meandering Caspian S connected Black S			Sidorchuck et al. (2001, 2003) Sidorchuck et al. (2008, 2009) Sidorchuck et al. (2011) Markova et al. (2019) A Panin et al. (2013) cit. Vandenberghie and Sidorciuk (2020) Matlakhova (2020)
22 a	Upper Don, Khoper River	19-17ka, first generation of large paleomeanders	15-13 ka, the second of macromeanders, activation of aeolian processes	Lowering of runoff		
23	River Kizilirmak, middle reach, Turkey	19 ka - incision, then 18 m aggradation, climate and subsidence LGM precipitation 1.9 times high than today;				Dogan (2010, 2011)
24	River Sakarya, Turkey	T3 (22m) - 41 - 30ka (aggradation)	LGM-H (30 - 9ka), incision 25 m, river formed large sinusoidal bend while incising through the bedrock	T2 (10 m) - 9ka...1.8ka aggradation		Gorur et al. (2001) Erturac and Kiyak(2017) Erturac et al. (2019) Erturac(2021)
25	Sakarya delta, Turkey				The marked truncation at the top at about 7- 8 ka during the latest lowstand of the Black Sea. Water depth of 105 m	Algan et al. (2002)
26	Kizilirmak delta				7.9ka - 6.8ka, a lagoon to pond environment with frequent connection to the sea	Berndt et al. (2018)
27	Kuban delta				7.4ka, age in base of sediment column of 20 m thick	Bolikhovskaia et al. (2017)

28	Dniester, Black Sea shelf	-20...-22 m, submerged archaeological sites, ages between 34 ka – 12 ka; buried by thick fluvial sediments					Kadurin et al. (2020)
29	Danube delta					9.8-9.5ka the Danube was building a ramp delta lobe above 40 mbsl. 8–7.5 ka early stage of delta development	Giosan et al. (2009) Vespremeanu-Stroe et al. (2017)
30	Danube Canyon, BS shelf edge	53-40 ka; -10...-20 m BS level (Surozhian), silty clay beds,	27-17ka; -100 m BS level	9.5-7.5 ka: -10 m BS level	7.5 – 3.5 ka; -5...-6 m BS level		Popescu et al. (2004; 2015) Caraivan et al. (2012; 2017) Fleitmann et al. (2009)
31	Sofular Cave, Turkey	A 50 kyr-long exceptionally well-dated and highly resolved stalagmite oxygen (d18O) and carbon (d13C) isotope record					Kern et al. (2018, 2019) Burstyn et al. (2019)
31a	Karaca Cave, Trabzon area, NE Turkey	An oxygen isotope record from a stalagmite that grew between 77 ka and 6 ka in northeast Turkey					Rowe et al. (2012)
32	Tăușoare Cave, Romania	Stalagmite from Tausoare Cave (TC), East Carpathians - data on past climate changes between 69 – 15 ka					Staubwasser et al. (2018)
33	Ascunsă Cave, Romania	A speleothem isotope record (POM2) that provides data on past climate changes in the S Carpathian region from 8.2 ka until the present. A speleothem (POM1) that provides data on past climate changes from 48 ka to 30 ka.					Drăgușin et al. (2014), Staubwasser et al. (2018)
34	Preluca Tiganului, peat, Romania	High-resolution pollen records providing data on past climate changes in the NW Carpathian area between 14.6 – 8 ka					Feurdean et al. (2014)

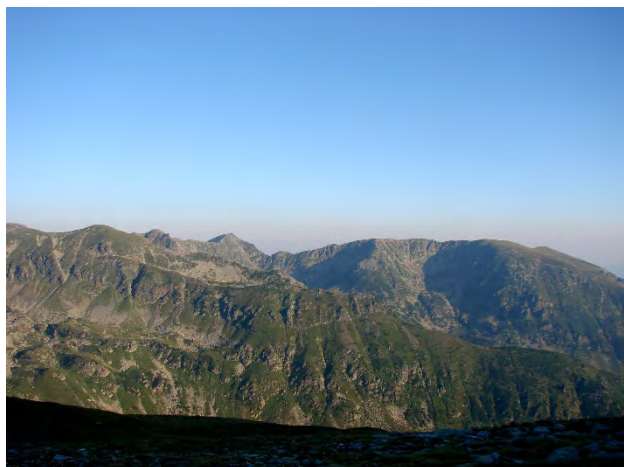
Symposium

National Symposium on Geomorphology, 36th Edition

Geomorphology, one step further

Craiova, Romania, September 22-26, 2021

The Romanian Association of Geomorphologists organized between September 22-26, 2021, the 36th Edition of the National Symposium of Geomorphology. After ten years, the Department of Geography of the Faculty of Sciences of the University of Craiova hosted for the second time this traditional scientific event that annually brings together geomorphologists from Romania and their guests.



This prestigious symposium of Romanian geography took place under the motto Geomorphology, a step towards the future, considering the special role in the society of the third millennium of the science of earth's crust landforms in terms of physiognomy, genesis and evolution over time.

This event was attended by geomorphologists from the Romanian academic space, the University of Bucharest, Aleandru Ioan Cuza University of Iași, Babeș-Bolyai University of Cluj-Napoca, West University of Timișoara, Institute of Geography of the Romanian Academy, University of Craiova, University of Oradea, together with guests from CNRS LETG - Nantes, France and the University of Liège, Belgium.



In the first part, the symposium papers were structured in four sessions, where the results of research focused on topics such as slope dynamics, river geomorphology, coastal geomorphology, geohazards, dendrogeomorphology, speleology, avalanches, etc. were presented. The four joint sessions (online) was a good opportunity to meet geomorphologists interested in connecting research with the needs of society.



The second part of the symposium was the field application in alpine space near Râncă Resort. The field application targeted the general

geomorphological context of Rânca resort, the implications of the modernization of DN67C (Transalpina) and the glacial and periglacial relief in the upper basins of Gilort, Lotru and Latorița, within the sculptural complexes Râu Șes and Borăscu.

With the hope that the geographers of the University of Craiova will have again the honor of hosting the National Geomorphology Symposium, and that the spirit of Oltenia "was a source of energy" and inspiration, we want to thank all participants.



Sandu BOENGIU, *Craiova University*

Symposium

International Geomorphology Week

1st-7th March 2021

The International Geomorphology Week is a worldwide event that aims to celebrate and promote geomorphology in all its forms. During the week, under the auspices of the International Association of Geomorphologists (IAG), were held online various geomorphology-themed activities to mobilize and gather geomorphologists around the world. Several free-to-attend regional webinars have been organized to facilitate and keep alive the connections within the geomorphological community. One other important aspect is that the webinars encouraged the young researchers to present their work and get in contact with their peers and their field-related specialists.

In numbers, the week summed 11 IAG Regional webinars: South & West Asia, Iberia, Central & Western Europe, Africa, Western Europe, British and Irish Isles, Northern Europe, Australia and New Zealand, South & Central America, Central-Eastern Europe, North America & Costa Rica, East and Southeast Asia; and more than ten other events such as the Young Geomorphologists meeting, student webinars, Mediterranean Women in Geomorphology (see more details at the following [link](#)).

The Romanian Young Geomorphologists had also presented their work within two of these webinars. In the Central & Eastern Europe IAG Webinar, **Florin Zăinescu** from the *University of Bucharest* talked about his work on the

Morphodynamics and flow interactions of deltaic wave-influenced river mouths. Two other scientific contributions were given during the IAG International Young Geomorphologists meeting. In the first one, **Laurențiu Țuțuianu** from the *University of Bucharest* presented the *Wetlands and lakes formation and evolution on the Lower Danube Floodplain during the Middle and Late Holocene*. In the second one, **Nicușor Necula** from the *Alexandru Ioan Cuza University of Iași* talked about the *Landslide dynamics assessment using Multi-temporal SAR Interferometry and numerical modeling*, followed up, in the second part of the conference, by the presentation of the Romanian Young Geomorphologists Group within the International Young Geomorphologists network.

While we face difficult times at the moment, online gatherings have become the norm which allows scientists to socialize and get in touch with each other, to fact-check, develop and discuss ideas in a similar face-to-face environment. The success of the International Geomorphology Week webinars had called for a follow-up meeting that is planned for 2022. See details at the following [link](#) and **Stay tuned!**

Representative of Romanian Young
Geomorphologists Group,
PhD Student Nicușor NECULA, *Alexandru Ioan Cuza
University of Iași*

Revista de GEOMORFOLOGIE

ARTICLES

Paolo BILLI - *Geomorphic characteristics of dryland distributive systems.*

Mihai NICULIȚĂ - *Geomorphological restitutions for the geomorphological regionalization of Romania: the Moldavian Plain case.*

Gabriel NANFACK, Etame SONE DIABE - *Évaluation de la susceptibilité des sols au ruisseau dans le bassin versant de la Knam (Ouest Cameroun), en appliquant la méthode Services de Conservation des Sols Numéro de Courbe.*

Nathalie ANNAVAÏ, Anselme WAKPONOU, Moussa IYA - *Les unités morphostructurales et pédologiques des monts Mandara face à la rétention en eau de surface.*

Răzvan POPESCU, Alfred VESPREMEANU-STROE, Olimpiu POP, Nicolae CRUCERU - *Rock glaciers vegetation colonization in Retezat Mountains: implications for morphodynamics and palaeoclimate reconstruction.*

Maria RĂDOANE - *A history of the circum-pontic river channels marked by climate and sea level changes during the Late Quaternary (25-8 ka BP).*

MISCELLANEA

- The 36th Romanian Symposium on Geomorphology, *Geomorphology, one step further*, Craiova, Romania, September 22-26, 2021 (Sandu BOENGIU)

- International Geomorphology Week, 1st-7th March, 2021 (Nicușor NECULA)

