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Identifying glacio–isostatic rebound processes using testate amoeba as palaeohydrological proxies; a case study from subarctic Québec, Canada

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ABSTRACT

Glacio–isostatic rebound is one of the most important landscape processes affecting the northern and northeastern coast of Canada and, therefore, reconstruction of postglacial rebound rates is critical for a better understanding of landscape evolution in this region. Yet, studies reconstructing coastal palaeogeography in Northern Canada are constrained by the limitation of shell–based radiocarbon chronologies used in dating shoreline displacement and palaeo–sea levels. This study proposes an alternative methodology for the reconstruction and dating of palaeo–sea levels, which uses testate amoeba microfossils from coastal sediments as palaeoecological markers of coastal water table shifts linked to sea level change. Our results indicate that testate amoeba–inferred water table reconstruction is a good indicator of water table drawdown due to isostatic uplift on affected coastlines. Furthermore, radiocarbon dating of distinct plant macrofossils within the transitional marine/freshwater stratigraphy avoids the inherent reservoir effect issues associated with dating of marine shells.

KEYWORDS

glacio–isostasy, wetland hydrology, paleoecology, testate amoebae, Transfer Function, radiocarbon dating

1. Introduction

The demise of the late Laurentide Ice Sheet during the early Holocene induced a large–scale process of isostatic rebound throughout northern and north-eastern Canada. In the James Bay region, the post-glacial isostatic rebound caused one of the highest

vertical shoreline displacement known during the Holocene – more than 300 m in 8000 years (Hillaire–Marcel, 1980). Despite this dramatic reorganisation of coastal geomorphology, it is still unclear whether the isostatic rebound was a major driver of land–

scape change beyond the immediate impact on coastal configuration. One study conducted in the Hudson Bay Lowlands (Glaser et al., 2004) – a region adjacent to James Bay – found that differential rates of isostatic uplift across the region was the principle driver of peatland development and that fast rates of uplift result in continuous alteration of peatland hydrology. As land emerges and marine influence retreats, shoreline displacement drives water level decrease across the region. Other studies, such as Pendea (2011) and Pendea and Chmura (2012) found that isostatic rebound had a major impact on coastal ecology, but inland areas were less affected.

Regardless of the type of study undertaken, the quality of the scientific data related to glacio-isostatic processes hinges, to a large degree, on precise dating and reconstruction of transitional marine–terrestrial stratigraphies. Most previous studies on palaeo–sea levels have used fossil marine mollusc shells, driftwood and palaeosoils to date and reconstruct the position of former coastlines (e.g., Allard and Seguin, 1985; Lajeunesse and Allard, 2003 a, b) or diatom–inferred salinity shifts linked to sea level change in coastal lakes (e.g., Saulnier–Talbot and Pienitz, 2001; Miousse et al., 2003, Fedje et al., 2018). Some of the limitations in these studies stem from the fact that marine shells are not an ideal material to use in radiocarbon dating due to the variable and difficult to constrain reservoir effect (e.g., Allard and Seguin, 1985; Lajeunesse and Allard, 2003 a, b). Moreover, the use of marine shells to reconstruct palaeo–sea levels is further complicated by the fact that many of the molluscs commonly dated in the James/Hudson Bay region have wide ecological tolerances with respect to water depth (Aitken and Fournier, 1993) and thus they may be poor proxies for the position of sea levels.

In studies where the marine/terrestrial stratigraphies were reconstructed based on diatom assemblage from former coastal lakes (e.g., Saulnier–Talbot and Pienitz, 2001; Miousse et al., 2003, Fedje et al., 2018), there is likely a lag in the decline of coastal lake water salinity that may vary with the lake volume and whether or not the system is open or closed (Pendea et al., 2010). This lag time between the reduction in lake salinity and the actual

local shoreline displacement can lead to an imprecise reconstruction of palaeo–sea levels.

In a more recent study, Pendea et al. (2010) used AMS radiocarbon dating of freshwater macrofossils from marine/terrestrial sedimentary transitions and analysis of pollen, spores, and dinoflagellate cysts to define the tidal stratigraphy – an ideal proxy for palaeo–sea levels. This study avoided many of the problems associated with imprecise dating of palaeo sea levels and demonstrated that palaeoecological proxies can be used to refine transitional stratigraphies.

In this study, we build on the findings of Pendea et al. (2010) and propose a methodology that provides direct evidence of water table drawdown due to isostatic uplift and better constrains the period of retreat of tidal influence. To this end, we use fossil testate amoebae assemblages from two sites in a depth–to–water table (DWT) transfer function to reconstruct coastal hydrology at the time of coastal emergence. We hypothesize that testate amoeba assemblages will indicate a sharp water table drawdown across the marine/terrestrial sedimentary interface as evidence for land uplift and retreat of tidal influence.

2. Study site

The study region (Fig. 1) is located in the high boreal zone of eastern James Bay, northwestern Québec (Canada). The region was covered by the last remnants of the Laurentide ice sheet until the early Holocene, after which marine waters from the Tyrell Sea began to flood the newly deglaciated surface (Dyke, 2004). The establishment of the marine stage was short, followed by shoreline regression under the influence of glacio–isostatic rebound. A progressive landscape emergence from east to west began ~7000 years ago in the study region and is currently still underway (Pendea et al., 2010). The James Bay regional climate is continental subarctic and is part of the Perhumid High Boreal Eco-climatic Region where the mean January temperature is -23°C , mean July temperature is 14°C , and the annual precipitation ranges between 528–833 mm (Crins et al., 2009). The regional vegetation consists of open boreal taiga, is dominated by stands of stunted *Picea mariana* with *Pinus banksiana*, *Alnus*

spp., *Larix laricina*, and *Betula glandulosa* as secondary constituents, and is interspersed with open fens and bogs (Crins et al., 2009).

Site W55 (53°1' N, 78°10' W) is a bog situated about 55 km east of the town of Wemindji (Québec). The site has an elevation of 130 m a.s.l. and is surrounded by relatively steep slopes towards the south. An intermittent brook on the northern margin partially drains the site, but the overall drainage

is poor due to the underlying marine clays. The peatland developed in a former cove on the north-west side of a newly emerged island after 6400 cal yr BP (Pendea et al., 2010). The peatland vegetation at W55 site is dominated by *Sphagnum* mosses and ericaceous shrubs with patches of sedges and lichens. Stunted trees (*Picea mariana*) border the margins of the bog.

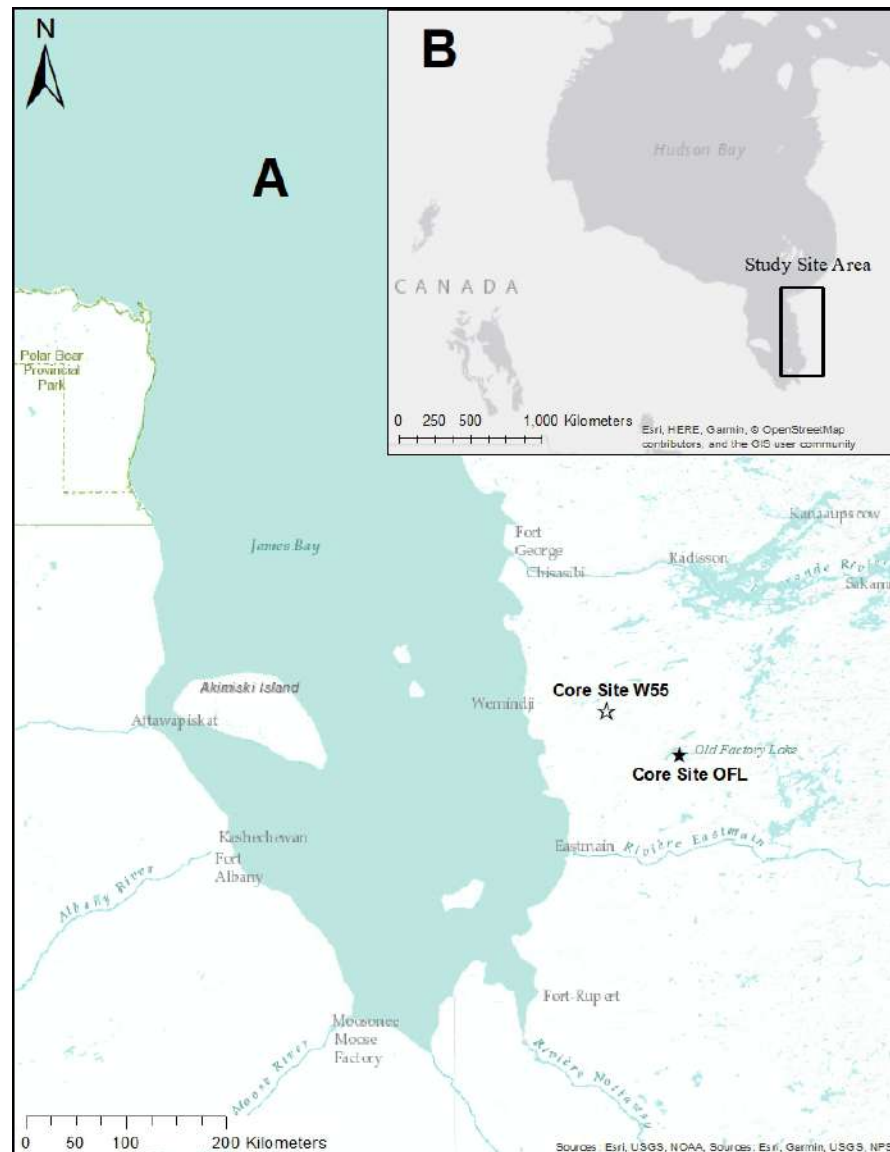


Figure 1 Location of the study region

Site OFL (52°47' N, 77°31' W) is a poor fen/bog complex situated on the southern shores of Old Factory Lake, approximately 110 km southeast from the town of Wemindji (Québec). The wetland has an

elevation of 103 m a.s.l and is characterized by a flat topography with no surface drainage. The surface vegetation is dominated by *Sphagnum* and ericaceous shrubs along with stunted *Picea mariana* trees

found isolated throughout the bog (Pendea, 2011). The peatland developed on a former tidal flat, which at the time of emergence was a large protected bay (Pendea et al., 2010). When the sea finally retreated, the former bay was replaced by the Old Factory Lake (Fig. 1). The local base level created by the Old Factory Lake is the primary control for the local water table level at this site. The subterranean drainage is very poor due to the thick marine clay layer at its base.

3. Materials and methods

3.1 Field and laboratory sampling

To reconstruct the marine/terrestrial transition and the palaeohydrology of the two sites, we used five-cm diameter cores extruded from the deepest part of each peatland with a 2 m-long Eijkelpamp® piston corer. The top 0.5 m of the cores, containing *Sphagnum* peat with tough ericaceous root systems was cut with a knife. Cores were transported in PVC pipes and stored in a cold room at 4°C until subsampled in the laboratory. Cores were then sliced into 0.5 or 1 cm-thick sections and from the center of each section a subsample was removed using a 1.5 cm-diameter metal cylinder. Thinner sections

were cut above and below the marine/terrestrial interface. Mean sampling resolution within the cores was 6 ± 3 cm (180 ± 166 yrs) for the W55 core, 6 ± 2 cm (133 ± 90 yrs) for OFL core. Core W55 had very low accumulation rates between 222 and 227 cm-depth.

3.2 Dating and age-depth modelling

Accelerated Mass Spectrometry (AMS) radiocarbon measurements were obtained exclusively on macrofossils including seeds, twigs, leaves, and *Sphagnum* stems and leaves (Table 1 and 2). AMS radiocarbon analysis was performed by Beta Analytic Inc. Quoted errors represent one relative standard deviation statistics (68% probability) and counting errors are based on the combined measurements of the sample, background, and modern reference standards. Radiocarbon ages were corrected for isotopic fractionation and were calibrated using the INTCal13 curve (Reimer et al., 2004) and the calculations were performed using the cubic spline fit (Talma and Vogel, 1993). Calibrated ranges are reported as two standard deviations (Tables 1 and 2).

Table 1 Radiocarbon measurements and associated calendar ages based on macrofossils from OFL core James Bay, Québec

Sample Depth (cm)	Dated material	Beta code	Radiocarbon age (yr BP)	2 σ probability CAL age (yr BP)		
				1st	2nd	Midpoint
204	<i>Sphagnum</i> stems and leaves	251801	3980 $\pm$ 40	4400–4530		4465
225	<i>Alnus</i> twig	251802	4040 $\pm$ 40	4770–4790	4420–4620	4780
233	<i>Alnus</i> bark and twig	251800	4320 $\pm$ 40	4970–4840		4905
237	Charred material (seeds, wood, and <i>Picea</i> needle)	257586	4970 $\pm$ 40	5830–5860	5600–5750	5675

Table 2 Radiocarbon measurements and associated calendar ages based on macrofossils from W55 core James Bay, Québec

Sample Depth (cm)	Dated material	Beta code	Radiocarbon age (yr BP)	2 σ probability CAL age (yr BP)		
				1st	2nd	Midpoint
208	<i>Alnus</i> twig	251807	3090 $\pm$ 40	3220–3390		3305
221	<i>Alnus</i> twig, <i>Betula</i> leaf	251806	3930 $\pm$ 40	4510–4480	4440–4250	4345
226	<i>Alnus</i> twig	251805	5630 $\pm$ 40	6490–6310		6400

Age–depth models for cores OFL and W55 (Figs. 2 and 3) were derived using the Bayesian Bchron model as incorporated in the Bchron R software package (Haslett and Parnell, 2008; Parnell et al., 2008). This model outperforms other age–depth models, as shown by a large–scale experiment conducted on Holocene lake sediments (Parnell et al., 2008).

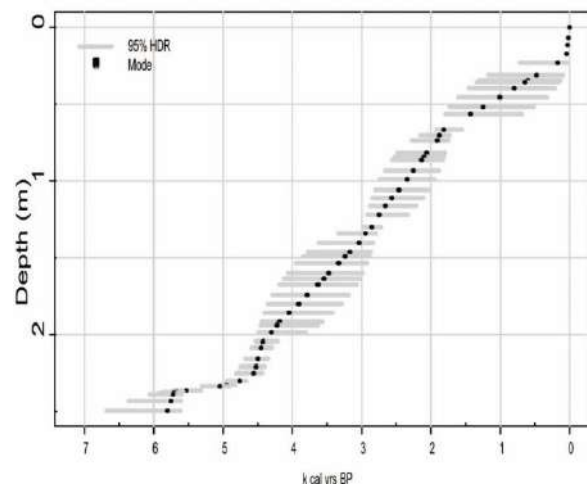


Figure 2 The age–depth model for core OFL based on radiocarbon dates as constructed by the Bchron software (Haslett and Parnell, 2008; Parnell et al., 2008). Grey bars represent the 95% confidence interval of calendar ages

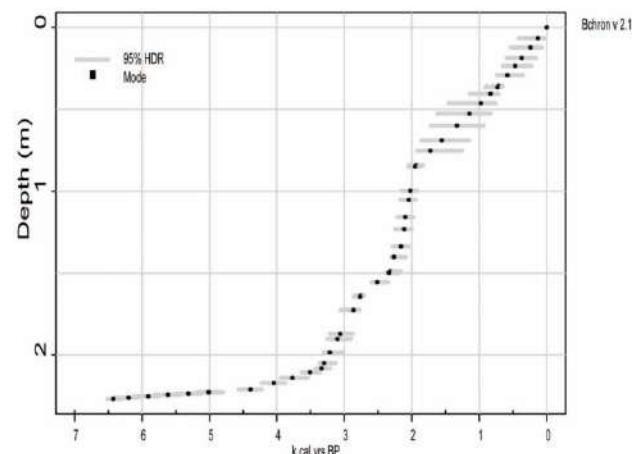


Figure 3 The age–depth model for core W55 based on radiocarbon dates as constructed by the Bchron software (Haslett and Parnell, 2008; Parnell et al., 2008). Grey bars represent the 95% confidence interval of calendar ages

3.3 Testate Amoebae sample preparation

Approximately 1.5 cm³ of peat material from the center of each 1–cm thick slice was put in test tubes where a *Lycopodium* tablet of known spore number

was added for calculation of test concentrations. The samples were then boiled in distilled water for 10 minutes followed by a two–part sieving process. The samples were passed through a 300–µm sieve and retained on a 15–µm sieve after thorough washing. Samples were then centrifuged to remove excess water and mounted on slides for identification. All samples were counted to 150 specimens to increase randomness and decrease possibility of missing a species. The taxonomic framework and nomenclature follows Charman (2001) as modified by Booth (2007). Taxa identification was conducted using a key published by Booth and Sullivan (2007).

3.4 Zonation

All taxa were used to subdivide the testate amoebae assemblage diagrams into local testate assemblage zones (LTAZ) using constrained cluster analysis by sum–of–squares (CONISS) as implemented in the Psimpoll 4.27 program (Bennett, 2007). Only zones representing significant clusters according to the broken stick model (Bennett, 2007) were retained and each is presented in detail in the Results section.

3.5 Depth–to–water–table transfer function

In this study, we used testate amoeba analysis to reconstruct the long–term hydrological variability for the OFL and W55 peatlands, measured as depth–to–water–table (DWT). Testate amoebae are protozoans that inhabit distinct ecological niches related to moisture conditions and produce morphologically distinct shells that can be used as palaeoecological indicators of hydrologic changes as they respond rapidly to changes in the hydrology (Mitchell et al., 2007).

These proxy indicators are particularly abundant in *Sphagnum* dominated peatlands. The distinct siliceous shells of testate amoebae preserve well in peat deposits making them complementary indicators to other long–established indicators such as pollen, spores and macrofossils (Mitchell et al., 2007). Authors such as Woodland (1998), Booth (2002, 2007), Booth and Zygmont (2005), Booth et al. (2010), Booth and Sullivan (2007), Amesbury et al. (2013), Lamarre et al. (2013), and van Bellen et al. (2014) have studied the ecology of testate amoebae

species and developed transfer functions from modern analogue data sets, which relate community species assemblages to mean annual water table depths. The transfer functions were then used in various peatland studies to infer current and past hydrology conditions.

Raw fossil test abundance data was transformed in percent relative abundance for each core. A transfer function was developed with the assistance of Dr. Robert K. Booth from Lehigh University using the weighted average partial least square (WA-PLS) model (Booth, 2008). The WA-PLS model returns

the smallest DWT errors although the general performance is similar to that of other commonly used models, such as weighted averaging (WA), weighted averaging with tolerance downweighting (WA-Tol), and weighted modern analogue technique (WMAT). The modern analogue dataset ($n=650$) is based on testate amoebae species–environment relationships from 369 sites across North America (Booth, 2008). The transfer function model was applied to our fossil dataset using the C2 software (Juggins, 2003; Booth, 2008) and mean water-table-depth was inferred with a mean error of 6–8 cm (Booth, 2008).

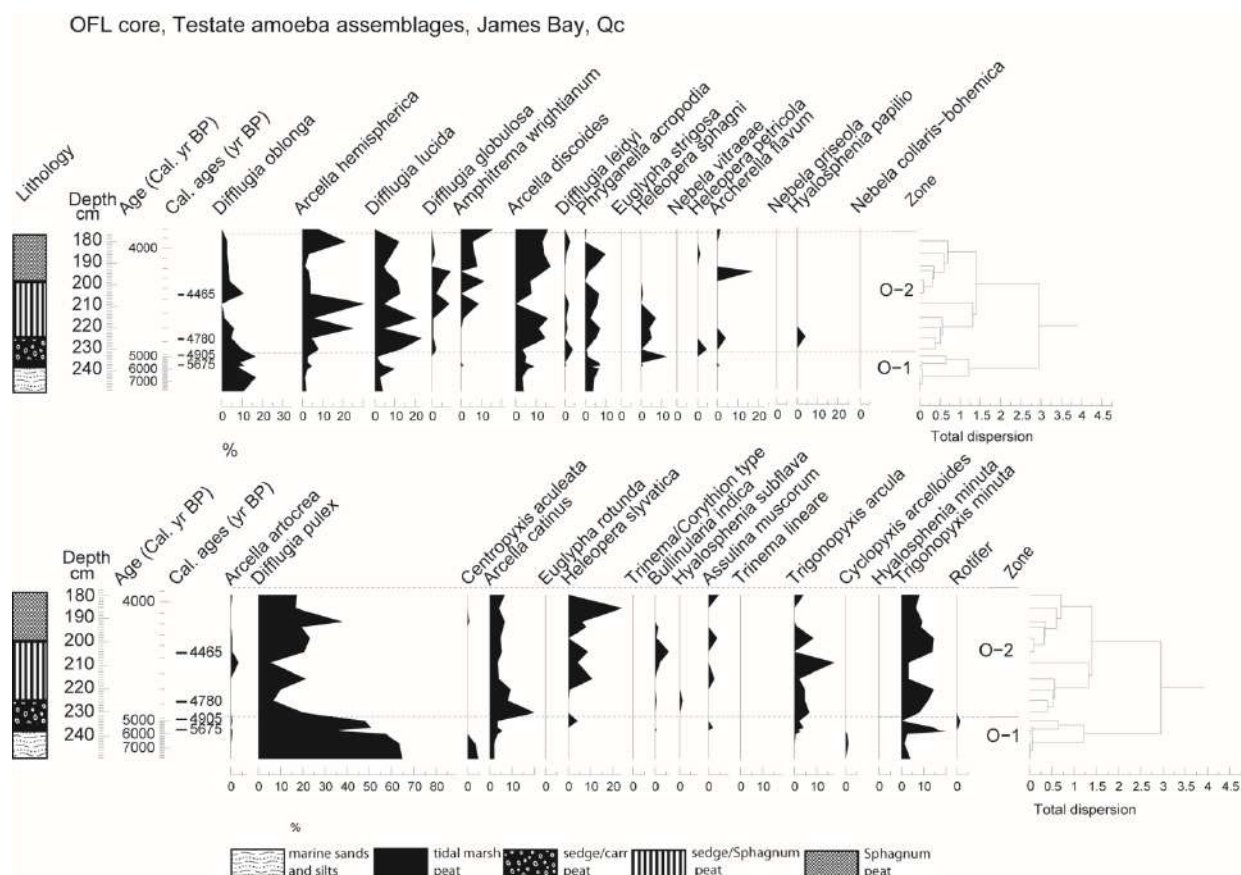


Figure 4 The fossil testate amoeba community representative for the transition between the marine and terrestrial environment at the OFL site, James Bay (Canada). The assemblage zonation was conducted on statistically significant clusters according to CONISS (Bennett, 2007)

4. Results

4.1 Chronostratigraphy

Table 1 and table 2 show the radiocarbon age measurements that circumscribe the stratigraphic transition between marine and freshwater environments. In addition to these age measurements, we used additional ages from Pendea (2011) to pro-

duce age–depth models for both the OFL and W55 core (Fig. 2 and 3). Although the additional age measurements used in the models are representative for the upper part of the cores that are not discussed in this paper, the age–depth modelling performs better when it uses age measurements across the entire core.

The earliest part of the OFL peatland development (5700 – 4900 cal yr BP) was characterized by low rate of sediment accumulation (Fig. 2). The model shows near constant accumulation rates between 4900 cal yr BP and present. For the W55 peatland, sedimentation rates were low between 6400 and 4000 cal yr BP, but increase significantly thereafter (Fig. 3).

4.2 Testate amoebae assemblages

We analysed 74 samples from the OFL core and 77 samples from the W55 core. A minimum of 150 specimens were counted in each sample, as suggested by Booth (2007). Testate amoeba species were identified to the lowest taxonomic level possible, resulting in 46 different species identified throughout the OFL peat core and 42 different taxa for the W55 peat core. A photo collection of some of the commonly observed testate amoebae species from both peat cores is presented in Plate 1. Percent relative abundance of testate amoebae was plotted against depth (cm) and age (cal yr BP) in Fig 4 and Fig 5, using Psimpoll 4.27 program (Bennett, 2007) for both the OFL and W55 peat core, respectively.

4.2.1. OFL core testate amoeba zonation

Zone O–1 (5900 – 4900 cal yr BP)

The O–1 zone (Fig. 4) consists mainly of *Diffflugia pulex*, which ranges from 40–65% abundance in this zone. Other important taxa are *Trigonopyxis minuta* (3–20%), *Diffflugia oblonga* (8–16%), *Diffflugia lucida* (2–10%), and *Heleopera sphagni* (0–10%). Several other taxa, including *Arcella artocrea*, *Centropyxis aculeata*, *Heleopera sylvatica*, *Bullinularia indica*, *Assulina muscorum*, *Trigonopyxis arcuata*, *Cyclopyxis arcelloides*, *Arcella hemispherica*, *Amphitrema wrightianum*, *Diffflugia leidy*, *Heleopera petricola*, and *Archerella flavum* are present in an abundance of 5 % or less.

Zone O–2 (4900 – 3900 cal yr BP)

Zone O–2 is characterized by a fluctuating abundance of *Diffflugia pulex*, with values ranging between 5 and 35%. Other important taxa are *Arcella catinus* (5–20% abundance), *Arcella emispherica* (5–30%), *Heleopera sylvatica* (0–25%), *Diffflugia lucida* (5–25%), *Arcella discoides* (0–20%) and *Trigonopyxis*

arcuata (0–20%). Taxa that are somewhat more irregular in frequency are *Archerella flavum* (1–20%), *Trigonopyxis minuta* (5–15%), *Amphitrema wrightianum* (8–10%), *Phryganella acropodia* (2–10%), and *Diffflugia oblonga* (2–10%). There are several minor taxa that display an abundance of 5% or less including *Bullinularia indica*, *Assulina muscorum*, *Diffflugia globulosa*, *Arcella artocrea*, *Diffflugia leidy*, *Heleopera petricola*, and *Hyalosphenia papilio*.

4.2.2. W55 testate amoebae zonation

Zone W–1 (6500 – 5000 cal yr BP)

The earliest zone of the W55 peat core (Fig 5) is characterized by the dominance of *Diffflugia pulex*, which reaches 20–50% abundance. Other significant taxa are *Diffflugia oblonga* (5–37%), *Archerella flavum* (5–35%), *Phryganella acropodia* (5–15%), *Heleopera sylvatica* (maximum 10%) and *Diffflugia globulosa* (maximum 10%). Minor taxa with abundance ranging below 5% are *Arcella catinus*, *Trinema corythion* type, *Hyalosphenia subflava*, *Trigonopyxis minuta*, *Arcella hemispherica*, *Diffflugia globulosa*, *Heleopera sphagni*, and *Nebela vitraeae*.

Zone W–2 (5000 – 4200 cal yr BP)

Zone W–2 is dominated by *Archerella flavum* with an abundance ranging from 48 to 65%. *Diffflugia pulex* decreases in abundance from 50% in Zone W–1 to 4% in W–2. Other important taxa are *Amphitrema wrightianum* (8–15% abundance), *Heleopera sylvatica* (0–10%), *Assulina muscorum* (3–10%), and *Diffflugia oblonga* (2–12%). Taxa that are minor (<5%) and/or infrequent are *Arcella artocrea*, *Arcella catinus*, *Trigonopyxis arcuata*, *Trigonopyxis minuta*, *Arcella hemispherica*, *Diffflugia lucida*, *Arcella discoides*, *Phryganella acropodia*, *Nebela vitraeae*, *Heleopera petricola*, and *Nebela griseola*.

4.3 Depth-to-Water table (DWT) reconstruction

4.3.1. The OFL peatland

The testate amoeba-based palaeohydrological reconstruction of the OFL peatland (Fig. 6) shows a DWT that is somewhat more variable than that of the W55 site. Prior to the stratigraphic transition from marine to terrestrial environments, the DWT was small (≈6–8 cm). During the shift from marine

clays to freshwater peat (5800 and 5600 cal yr BP) the DWT shows a marked increase, amounting to a water table drawdown of up to 20 cm. This surface

drying was brief and the peat surface returned to wetter conditions after 5500 cal yr BP.

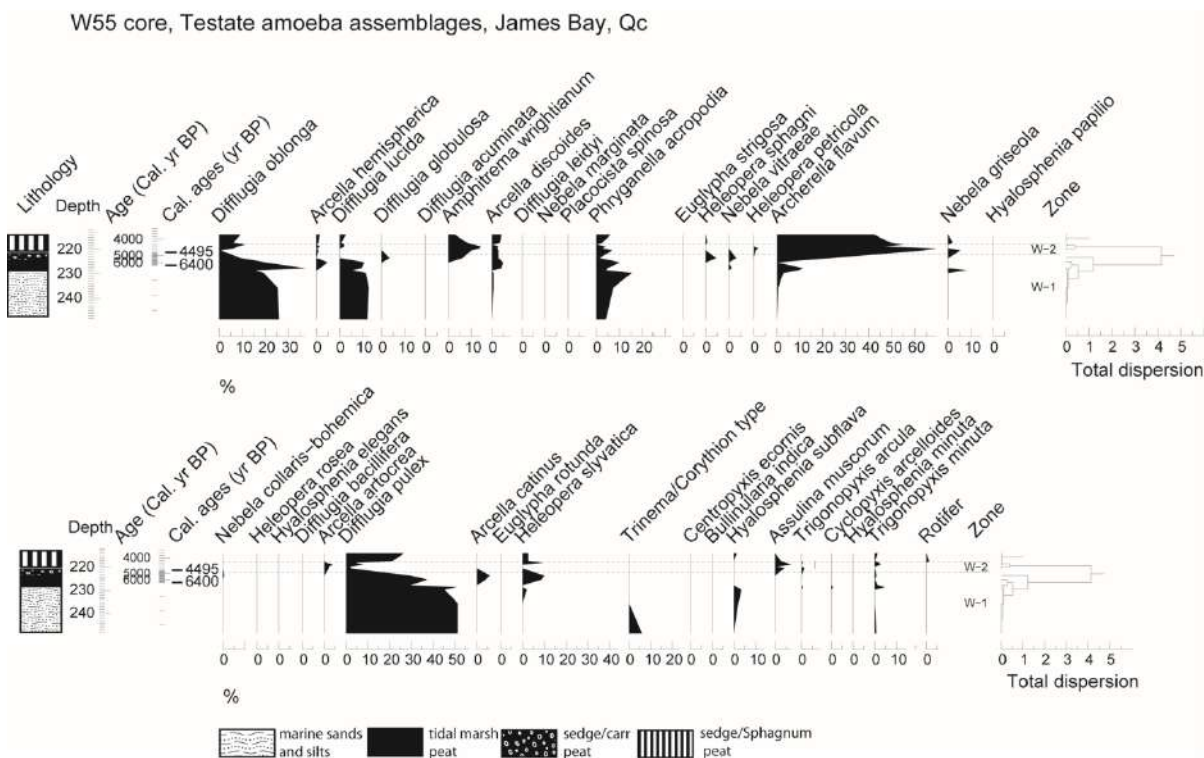


Figure 5 The fossil testate amoeba community representative for the transition between the marine and terrestrial environment at the OFL site, James Bay (Canada). The assemblage zonation was conducted on statistically significant clusters according to CONISS (Bennett, 2007)

4.3.2. The W55 peatland

The palaeohydrological reconstruction of the W55 peatland (Fig. 7) shows a less fluctuating DWT pattern than the OFL core. During the marine and tidal stages of this site (prior to 6450 cal yr BP), the testate amoeba assemblages indicate, as expected, high water tables with water pooling at the surface. The testate amoeba assemblages retrieved from tidal to freshwater transitional sediments (6450 – 6200 cal yr BP) indicate a marked but brief water table drawdown at the site. The DWT increases again after 6000 cal yr BP and remains relatively stable during the early freshwater phase of the site.

5. Discussion

The utility of testate amoeba as tracers of glacio-isostatic rebound processes

Glacio-isostatic processes are large-scale vertical movement of tectonic plates in response to glacial

loading and unloading. The most dramatic effects of these processes are illustrated by shoreline displacement in former glacial regions that underwent marine transgressions. In the James Bay region of northeastern Canada, the postglacial vertical isostatic adjustment totals over 300 m and this process was marked by a continuous marine regression that is still underway. The importance of this shoreline displacement cannot be understated as ecosystems and people have been continuously affected by dramatic change (Pendea et al., 2010; Pendea, 2011). It follows therefore that our ability to understand these large-scale landscape changes and their impact in the future depends, to a large degree, on precise reconstruction on timing and rate of change of shoreline displacement.

Research on isostatic coastline has concentrated on two aspects. The first aspect is the reconstruction of isolation stratigraphy, that is, the sedimentary sequence associated with the transition from marine

to terrestrial environments. The second research focus is the precise dating of the isolation event, defined as the period of time when the newly emerged landscape was isolated from the marine waters. Both these research objectives require proxy methodologies that are able to reconstruct events long past.

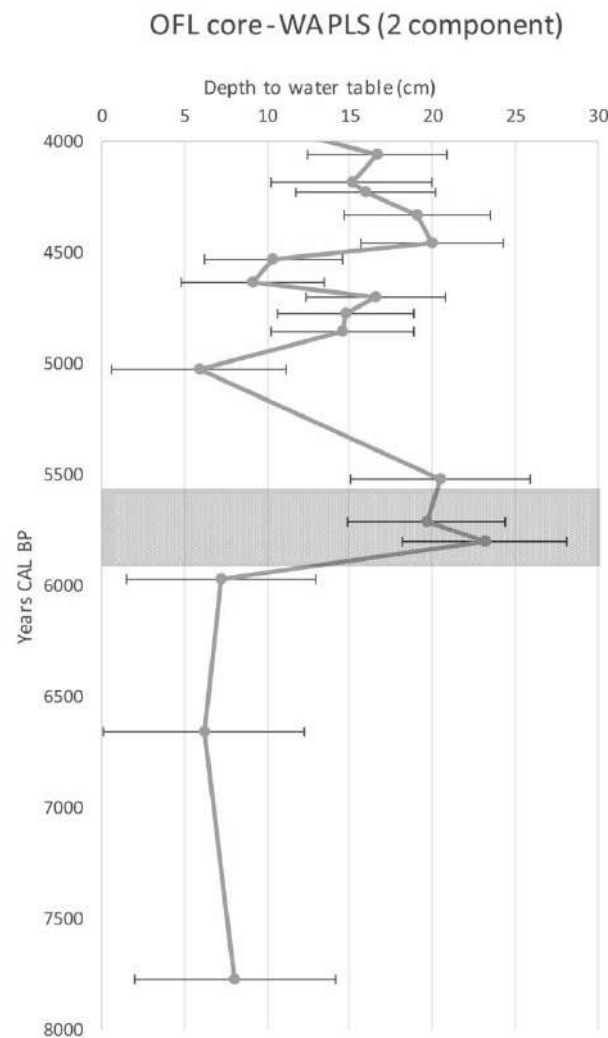


Figure 6 Testate-amoeba inferred water table depths for the OFL core (James Bay, Canada) using the WA-PLS model transfer function. The shaded interval represents the stratigraphic transition between marine and freshwater environments

In Canada, many studies used fossil marine mollusc shells to determine the position of the coastline at any given time and radiocarbon dating of the same shells to determine the shoreline displacement chronology (e.g., Hardy, 1977; Allard and Seguin, 1985; Lajeunesse and Allard, 2003 a, b). The main

problem with this approach is related to the relatively large potential errors in reconstruction and dating of palaeo-sea levels.

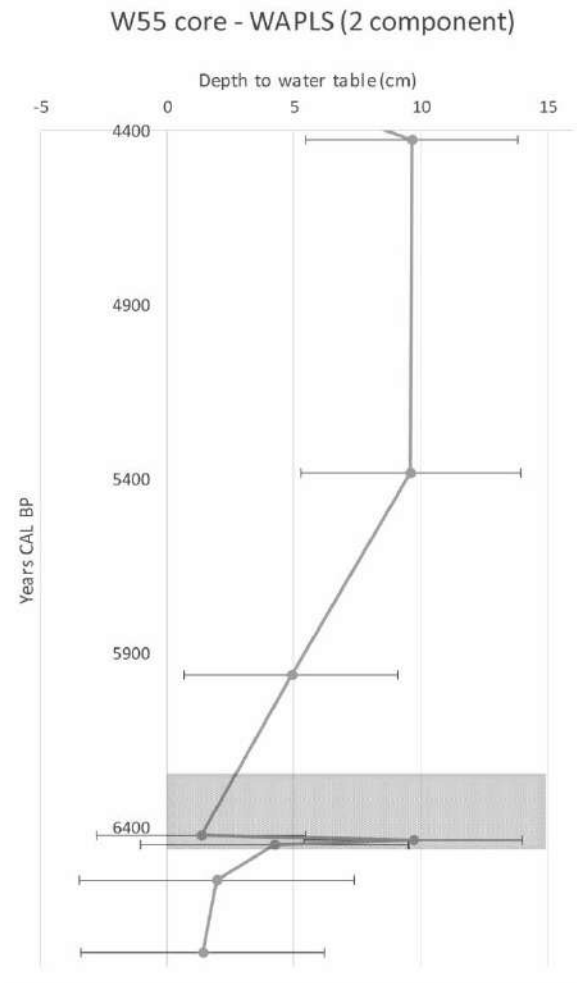


Figure 7 Testate-amoeba inferred water table depths for the W55 core (James Bay, Canada) using the WA-PLS model transfer function. The shaded interval represents the stratigraphic transition between marine and freshwater environments

Mollusc species that are commonly dated in the James Bay region, such as *Macoma calcaria*, can be found in shallow waters but also has been trawled from depths as great as 300 m on the continental shelf of Baffin Island (Aitken and Fournier, 1993). In such a case, the potential error in determining the vertical position of a coastline can be as high as 300 m. These shells have also been radiocarbon dated to provide timing for the reconstructed sea level curves. However, the inorganic carbon present in these shells can severely overestimate the actual

age of the shells, which in James Bay can be by more than 1000 years (Dyke et al., 2003; McNeely et al., 2006). Other studies such as Saulnier–Talbot and Pienitz, (2001), Miousse et al. (2003), and Fedje et al. (2018) have used diatom assemblages to reconstruct shifts in salinity related to the transformation of bays into coastal lakes and dating of organic remains associated with these shifts. This approach represents a major improvement as it avoids the large chronological errors associated with the reservoir effects in mollusc shells. However, the diatom–inferred salinity shifts are only representative for the isolation event if the residual salinity left in the newly formed lakes is immediately eliminated through drainage. If that is not the case, the marine–specific diatom assemblage will continue to subsist in the sedimentary record and thus underestimate the age of the isolation event.

A third approach was proposed by Pendea et al. (2010), who used pollen, spores and marine plankton fossils to reconstruct tidal environments as representative for the mean position of the coastlines. They also dated plant macrofossils from these environments, which were not affected by the reservoir effect. Although this methodology avoids many of the issues characterizing the other two approaches, it brings in a new challenge related to pollen taphonomy and dispersal. Pollen and spores preservation in sediments depends of the fossilisation environment and some pollen types do not preserve well in environments with frequent change in moisture such as tidal environments. Pollen and spores can also be dispersed over a large area and thus it can be difficult to relate a certain pollen assemblage with a specific location.

In this study, we proposed the use of testate amoeba as markers of change to soil hydrology due to isostatic uplift. The use of testate amoeba in sea level studies has already been demonstrated by Gehrels et al. (2001) in UK saltmarshes, who quantified species preferences for various levels of tidal flooding. However, most research agree that testate amoeba are excellent indicators of soil hydrological regime because assemblages respond very quickly to changes in water table depth (Mitchell et al., 2007). Indeed, our results show a sharp fall in water

table depth within the marine/terrestrial stratigraphic sequence most probably related to land uplift and subsequent seaward soil and surface drainage. The unambiguous record of water table drawdown in both cores corresponds well with the palaeoecological record of Pendea et al. (2010), who used pollen, spores and dinoflagellate cysts to reconstruct coastal emergence at the same sites. In their study, however, the reconstructed emergence interval is longer perhaps due to the fact that the palaeoecological proxies used to pinpoint emergence are indicative of the resulting ecosystems rather than the physical retreat of water from any given location. In this sense, testate amoeba help further constrain the emergence time and thus increase the precision of reconstructed sea level curves.

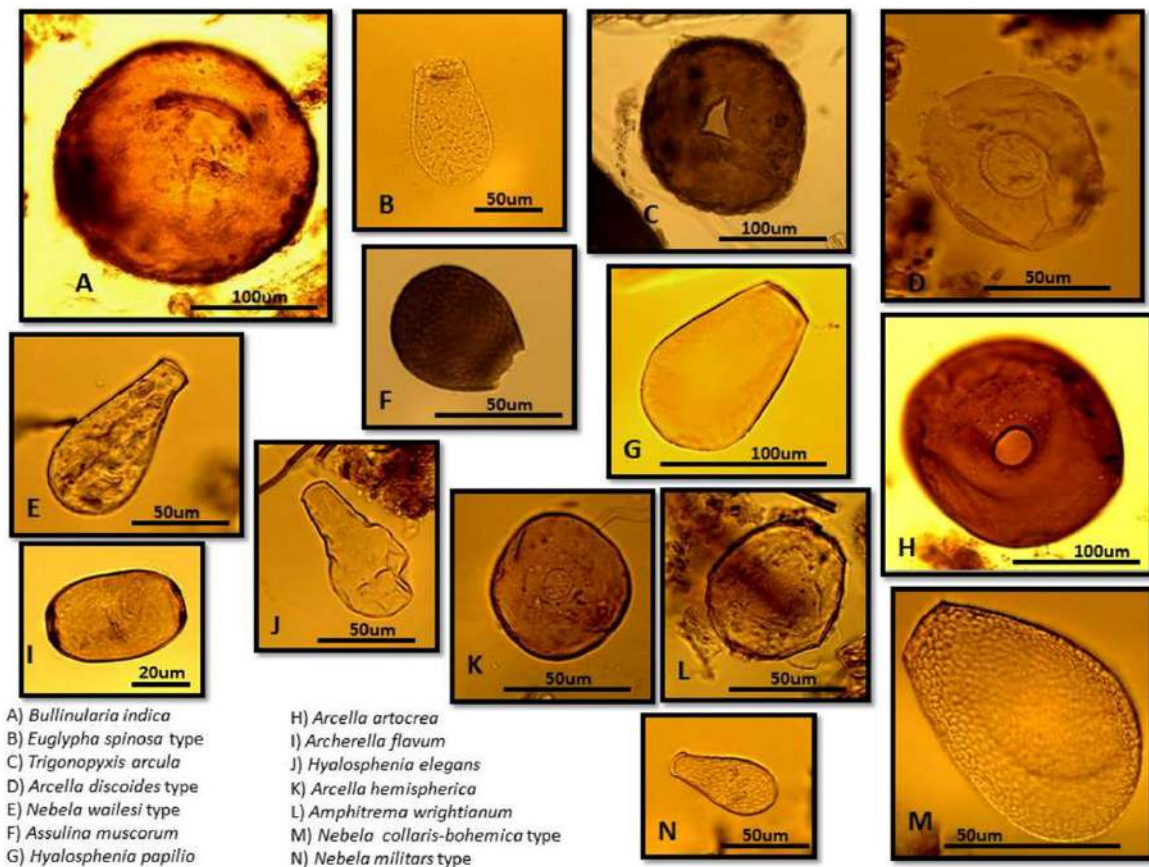
6. Conclusions

Our study demonstrates the potential for testate amoeba analysis in tracing coastal change related to isostatic rebound. Testate amoeba complement other palaeoecological indicators used previously (e.g., pollen, spores, marine and freshwater plankton), but may be more sensitive to the actual sea level fall by recording the local water table drawdown. However, future studies need to concentrate on calibrating the testate amoeba's ecological responses to sea level change via other parameters (e.g., salinity, hydroperiod). As Mitchel et al. (2008) pointed out, the reliability of this method depends to a large degree to a better understanding of taphonomy and ecological affinities of these microfossils.

Acknowledgements

We would like to thank Dr. Robert K. Booth from Lehigh University for his valuable assistance with the use of DWT transfer function for testate amoebae in peatlands of North America. We are also indebted to the Wemindji Cree First Nation for their support and fieldwork assistance. F. Pendea would like to thank fellow fieldworker Jesse Sayles for his friendship and assistance during the fieldwork season of 2007.

Plate 1 Photo collection of commonly observed testate amoebae for OFL and W55 peat cores, James Bay, Québec



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Mid and Late Holocene flooding reconstruction based on fluvial archives of the East Carpathian rivers

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ABSTRACT

This paper focused on the analysis of the interpretation potential given by fluvial archives, in order to decipher the Mid- to Late Holocene sedimentary history. The database discussed here refers to 65 dated ages (mostly radio-carbon), mappings of the most representative floodplain reaches of Suceava, Moldova and Siret rivers, and 23 outcrops analysed in floodplain deposits (river banks or gravel pits). The study results concentrated on (a) describing the regional features of Eastern Carpathian floodplains, (b) the floodplain morphology and its correlation with the structure of fluvial deposits and (c) identifying the paleoevents of fluvial activity on a centennial timescale. In this way, we were able to identify a general pattern of the dynamic geomorphology history for the rivers situated eastward of the Carpathians, which resulted from combining the Mid- to Late Holocene sedimentation phases, the avulsion and lateral migration processes that occurred during high fluvial activity intervals and the morphological adjustment of the rivers channels, from braiding to wandering and sinuously-meandering. The chronological overlay of the events revealed that the time intervals with the most extensive fluvial processes on the rivers draining the Eastern Carpathians occurred (in years before present) during 6200, 5300, 4100–4080, 3600, 2300–2200 (Iron Age), 1300 (Migrations Period), 950–970 (Medieval Climate Anomaly), 750–520 BP (Little Ice Age) and that these intervals generally overlap the high precipitation palaeoclimatic episodes documented in Romania. The fluvial activity of the rivers during the Mid- to Late Holocene has been compared as well with several records of paleo-climates variability over the geographical area of Romania. The result showed that many flooding episodes were coincident with the wet and cold intervals which were also identified in the variation of the other proxy data, especially after 4.7 kcal BP.

KEYWORDS

fluvial sediments; age dating; channel sedimentation and incision; flooding events chronology

1. Introduction

The paleohydrology and fluvial paleogeomorphology studies in Romania are relatively scarce. It was only after 2010 that a real recovery manifested in this way, the Holocene paleoevolution getting the attention of geomorphologists within a broader spectrum of landform types, not only with respect to the fluvial one. Various analyses starting from the integrated reconstruction of the Holocene evolution of the Danube Delta based on new evidences (Vespremeanu–Stroe et al., 2017), to the estimation of landslides activity during the Holocene on the sites with archaeological inheritance from the Moldavian Plateau (Niculiță et al., 2016) or to the evolution of sedimentation in small lake bodies (Mîndrescu et al., 2017) represent the range of references that confirm the statement above. The paleoclimatology studies complete these researches by contributing to understanding the role of climate in Holocene landforms dynamics. The synthesis of Perșoiu and Rădoane (2017) reviews the most notable studies regarding the knowledge of river channels patterns and change degree at different time levels (centennial and millennial, Late Holocene, Holocene, Late Glacial and the transition to Holocene). The most important conclusion of this synthesis reveals the high sensitivity degree of the rivers response to changing control factors, whether it is climate change, base level modification or human impact. The archive that preserves the “witness of change” status (according to Cordier et al., 2017) is built of alluvial assemblages created by authigenic sedimentation processes in a floodplain sector (Lewin et al., 2005). Despite their numerous limitations (listed by Lewin and Macklin, 2003), fluvial archives are considered valuable paleoenvironmental records, as proved by the publications of the FLAG work group (Fluvial Archives Group, 1996), a component group of the Water Framework Directive (WFD, 2000) of the European Union.

In this study, we aim to highlight the progresses in the interpretation of alluvial facies in correlation to floodplains morphology to ultimately discern the hydro-geomorphological paleoevents. This is an idea that started from the study of present river channels dynamics and of the human intervention

influence in the last 50 years (Rădoane et al., 2013), continued with comprehending the mechanisms that adjusted the fluvial patterns for the last 150 years (Rădoane et al., 2017) and further on, for the Mid to Late Holocene (Chiriloei et al., 2012; Rădoane et al., 2015) and Holocene (Rădoane et al., 2018). Here, we intend to harness our experience and database concerning the fluvial sediments archives from rivers floodplains and to illustrate this on a number of study cases of rivers from the eastern area of Romania. Amongst the discussed problems, we will particularly refer to: a) regional characteristics of the East-Carpathian floodplains; b) floodplains morphology and its correlation with floodplain deposits structure; c) deciphering the paleoevents of fluvial activity by using absolute dating of the fluvial sedimentary units.

2. Geographical setting

Our researches are located on multiple reaches of three important rivers that drain the eastern side of the Eastern Carpathians: Siret river and its right tributaries, Suceava and Moldova rivers (Table 1, Fig. 1).

The geologic substrate of Siret river basin is formed by a succession of strips oriented NW–SE that unfolds from west to east as follows: the volcanic area (1.33% of the basin area), the crystalline–Mesozoic area (6.79% of the basin area), the Flysch area (33.29%), the Neogene molasse area (10.12%) and the area of the platform-like sedimentary rocks (47.94%). The relief shaped on this substrate develops in steps from west to east and it is fragmented by a hydrographical network with an asymmetric layout towards the main river, Siret. The development of a well-individualized piedmont at the eastern border of the Carpathians and Subcarpathians is a relevant indicator of the long-term existence of coarse alluvial sediments consistent transport. This specific area is well-known to the Romanian geomorphologists as the Moldavian Piedmont, which started to form in the Sarmatian (on Fig. 2, this area is highlighted by the greater thickness of the Quaternary deposits from valley bottoms). Presently, this piedmont-like area is partly eroded, especially in its northern sector.

Table 1 Data on the studied river reaches

River	Drainage basin area upstream of the study area, A (km ²)	Reach length, L, (km)	Average channel slope, S, (m/km)	Discharge, mean annual, Q*, (m ³ /s)	Discharge, annual maximum, Qmax*, (m ³ /s)	Suspended load, mean annual, Qs*, (kg/s)	Bed material median diameter, D50, (mm)**
Siret (between Suceava cf. to Moldova cf.)	5,921	144	0.41	36.80 ^a	503 ^a	52.9 ^a	11.2
Suceava (between Milișăuți to Siret cf.)	2,334	128	2.00	16.49 ^b	381 ^b	21.9 ^b	20.0
Moldova (between Mold to Siret cf.)	4,316	110	0.82	32.80 ^c	546 ^c	35.3 ^c	17.0

*measurement period: 1967–2010; a – Lespezi gauge station; b – Țicani gauge station; c – Roman gauge station; ** cf Rădoane et al., (2008)

The piedmont-like activity is still present and it is visible in the behaviour of the river channels when leaving the mountain area (for example the braiding and even the avulsion and longitudinal profile deformation of Siret river – Ichim and Rădoane, 1990; Rădoane et al., 2008). On the investigated sectors, the rivers are semi-constrained (with a constraint degree of 35–40%), more significant on Siret river because of the contact with the Moldavian Plateau and with the front of the high terraces. Differences occur between the current typology of Siret river channel (sinuously meandering) and the one of the two tributaries (braided channel with tendency towards wandering near the confluence with Siret, more sharp in the case of Suceava river). Instead, a relative balance is noticed between the studied river reaches in terms of dimension, the type of material in the channel, the amount of water and sediment discharges (Table 1).

3. Data and Methods

The data providing the basis of the results presented here were obtained from our own researches comprising extensive field activities at 23 sites distributed along Siret (6 sites), Suceava (7 sites) and Moldova river (10 sites) which consisted of anthropic outcrops (sand and gravel pits) and natural exposures (river banks) in floodplain deposits. Radi-

ocarbon and OSL dates were published (Chiriloaei et al., 2012; Rădoane et al., 2015) or are currently in press in the framework of country-scale syntheses (Rădoane et al., submitted).

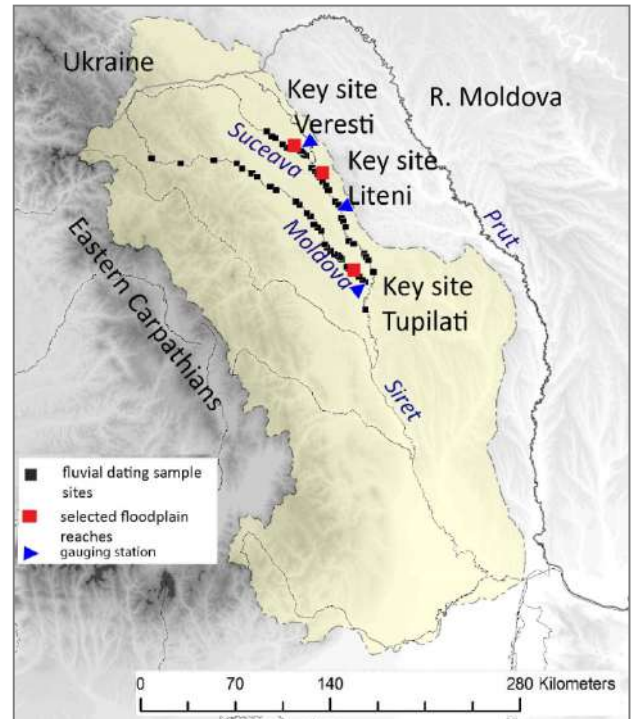


Figure 1 Location of sites along the studied rivers, with 14C dated river flood sediments. For each river a key floodplain reach was selected

The absolute ages database is made of 65 age dates of the fluvial sedimentary structures (60 radio-carbon, 2 OSL (Optically Stimulated Luminescence) and 3 dendrochronological datings; the latter dating method was applied to sub-fossil oak trunks trapped in the floodplain sedimentary structures).

For each dating, the following additional details were recorded: geographical setting (relief unit, drainage basin, river name, and geographical coordinates), altitude, distance from the source of the river at each site, the drainage basin area upstream the site location, information regarding the collected material, the stratigraphic position and the depth of the dated formation reported to the floodplain level. The spatial distribution of these data is presented in Fig. 1. Based on the experience accumulated in cautiously selecting age dating (Macklin and Lewin, 2003; Johnstone et al., 2006; Benito et al., 2015; Rossato et al., 2015; Faust and Wolf, 2017), in each distinct case we recorded the sedimentation context from which the sample had been selected. Every outcrop has been mapped and drawn to scale in order to identify and describe the sedimentary facies and the architectural elements such as in Miall (1985). The database obtained in this way was integrated in the general context of East-Carpathian rivers floodplains formation, the general characteristics of each being discussed below.

Therefore, the 14C, OSL and dendrochronological age datings are distributed on an altitudinal range between 200 and 550 m. These are the altitudes that comprise the floodplains extent of the three investigated rivers where we encountered frequent outcrops and datable samples. The greatest number of age datings ranges in the altitudinal class between 250–300 m (Fig. 2A). The drainage basin areas controlling the dated floodplain sections extend between 2000–7000 km² (in the case of Suceava and Moldova rivers) and between 9000–12,000 km² for Siret river. The jump in the size of drainage basins is visible in Fig. 2B and it is given by the fact that Suceava and Moldova are VIIth order rivers (in Strahler system) compared to Siret, which is a VIIIth order river.

The calibrated ages of the fluvial sedimentary units indicate a plethora covering the last 10,000 years (Fig. 2C), with a higher density in the last 7000

years. Their distribution with the sampling depth presents a wide variation of the fluvial units' age on the same depth, but also a slight age increasing tendency when sample location are getting deeper. In other words, as the sedimentary unit is located deeper, it gets older. On the other hand, the correlation is quite low-sensitive, due to the fact that fluvial sedimentation in floodplains undergoes intense and frequent reworking, therefore datable samples from older sediments could be eroded and re-deposited in a stratigraphic position closer to the surface. Nevertheless, the general tendency of the oldest sedimentary structures being found at the highest depths in the floodplain deposits is noteworthy.

The investigated fluvial sedimentary units are indicated in Fig. 2D, where they were classified as proposed by Miall (1985) and Lewin et al. (2005). The highest share of the identified and dated sedimentary structures belongs to channel bed gravel deposits opened in the channel banks and mostly in gravel quarries. In such deposits we identified the trunks of the subfossil trees that we dated (cf. Figs. 4, 5, 6). Some other categories of fluvial units were the gravel lobes with finer grain-size due to lateral accretion, the fine gravel and sand bars, paleo-channels fillings and flood deposits, all of which were identified in the openings from the lower floodplain adjacent to the river channel, or in the floodplain terraces.

In the following, we will insist on several defining characteristics of the East-Carpathians rivers based on the hundreds of drillings performed, in order to better comprehend the specificity of the Holocene and recent fluvial activity of these rivers.

4. Some characteristics of the East-Carpathian rivers floodplains

The general features of the East-Carpathian river valleys, especially those located north of Trotuș fault, reflect their evolution on nearly the same courses ever since the Sarmatian (~13.5 million years ago) (Donisă, 1968; Rădoane et al., 2003). The energy-environment of rivers maintained high for a long time, leading to the transport of large amounts of coarse sedimentary materials for a longer period of time. This transport and sedimentation pattern is

reflected by the structure of the East–Carpathian rivers fluvial terraces and especially by their floodplains. The alluvial history of Siret river represents a distinct situation. Although this river flows mostly over a plateau, it has all the features of a Carpathian river, because of the granulometric and petrographic natures of the deposits from the river bed. The large amount of coarse sediments brought by the right bank tributaries from the Carpathians, caused much aggradation of the river bed and consequently, distorted the longitudinal profiles (Rădoane et al., 2008).

The floodplains of the East–Carpathian rivers are characterized by a morphology and structure of the

sediments deposit that reflects the Late Quaternary fluvial evolution common to all the rivers from Central and Eastern Europe (cf. Starkel et al., 2015; Perşoiu et al., 2017). The oldest formations identified and dated so far covered mainly the Middle Pleniglacial – Late Glacial period (MIS 3, MIS 2; ca. 59,000–11,700 yrs BP). The rivers that drain the SE Carpathians provide an exceptional situation, as the positive tectonic movements have determined a 250 m incision of the river network. The age of the fluvial formations that accompany these valleys was documented so far as the Eemian – Early Pleniglacial (MIS 5, MIS4) (Necea et al., 2013).

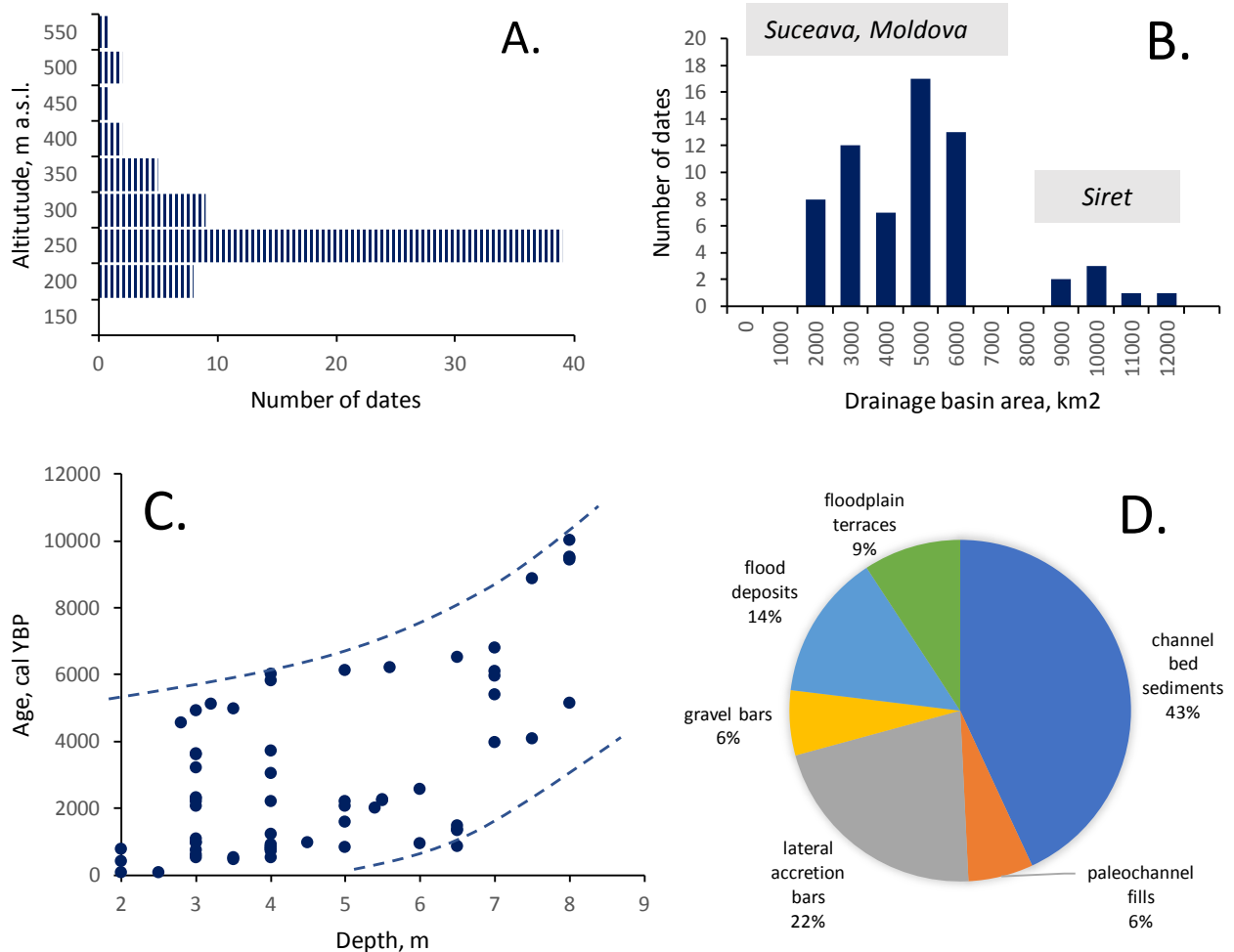


Figure 2 The distribution of age datings of the studied fluvial archives: (A) number of age datings of the investigated sites on altitude classes; (B) frequency of datings on size classes of the drainage basins situated upstream of the study sites; (C) the age of the fluvial units in relation to sampling depth; (D) the distribution of age datings by type of fluvial deposits

The several hundred boreholes performed in the floodplains of Siret river and of its tributaries coming from the Carpathians (Suceava, Moldova) (Fig. 3. A, B) have shown that the longitudinal profile of the fossil river-bed is not regular, presenting an alternation of deepening sectors (up to 18–20 m in this case), separated by sectors in which the fossil

bed is found at a lower depth (4–8 m). On Suceava river, the greatest thickness of the floodplain deposits occurs at the contact between orogeny and platform, overlaying the Rădăuți depression where there is a considerable widening of the valley bottom (Fig. 3A).

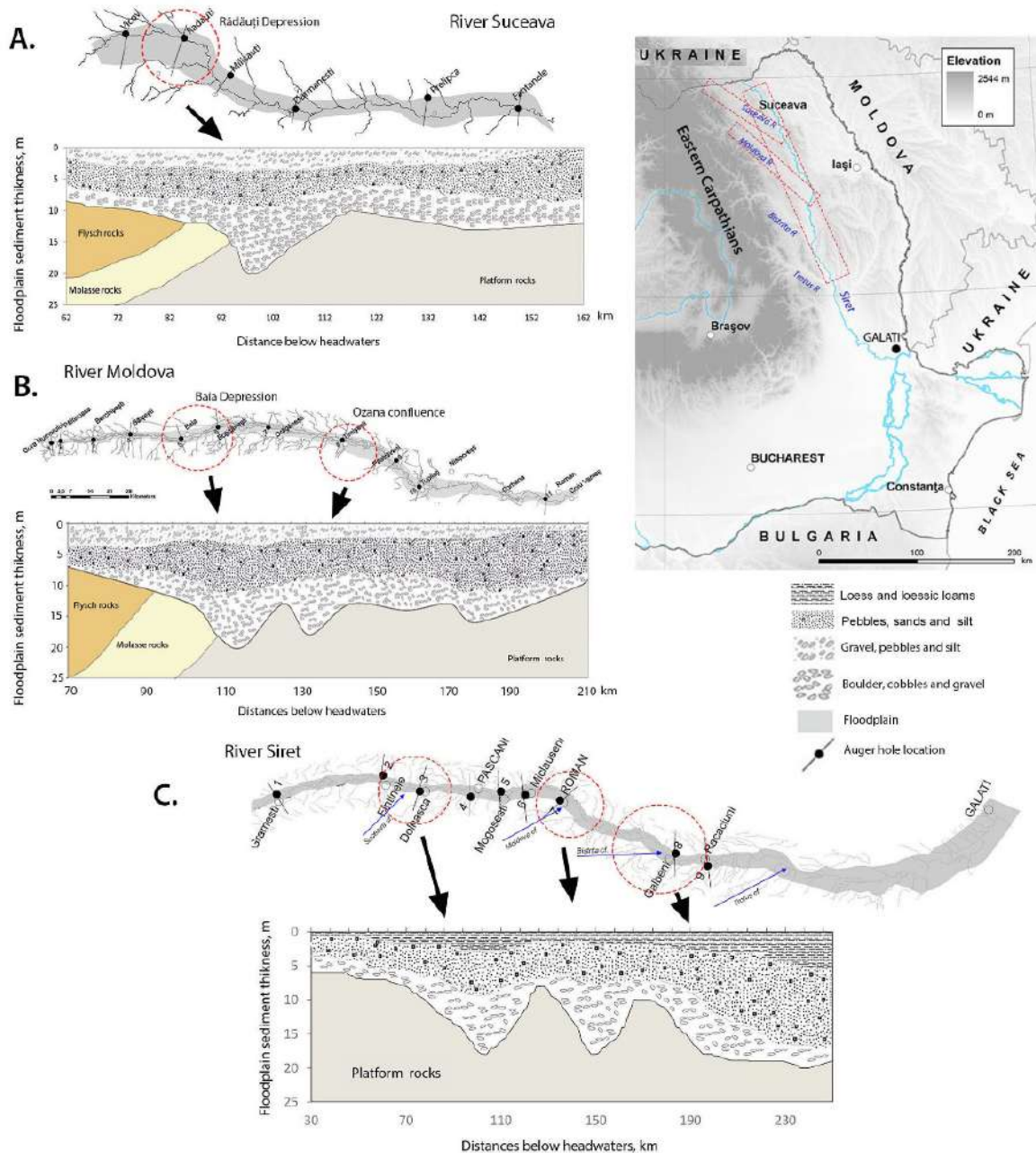


Figure 3 Suballuvial morphology and sedimentary fill of some Carpathian rivers: A. River Suceava, middle reach; B. River Moldova, out – Carpathian reach; C. River Siret, out – Carpathian reach; D. Exemplified rivers and their location in the geographical area of Romania. Discussions in the text

A similar phenomenon can be noticed on Moldova river, at the exit from the mountainous area of the river (Baia depression marking the area at the surface), where the thickness of the alluvial sediments reaches maximum depths. Additionally, the depth of the alluvial sediments increases along the valley at the confluences with prolific tributaries such as Ozana or Topolița. In the case of Siret floodplain, the suballuvial morphology is strongly controlled by its forceful Carpathian tributaries, so that at the confluences with Suceava, Moldova, Bistrița and Trotuș the bed of the floodplain decreases with 15 to 20 m (Fig. 3C).

Regarding the structure of the floodplain alluvial deposits, the boreholes allowed us to synthesize the stratigraphic arrangement of the valleys filling deposits, as follows:

i) a thick complex, formed by gravel and boulders, generally situated in the alluvial bed of the valley on all the three exemplified rivers. Its thickness varied from 4–5 m in the deepening sectors of the fossil bed and up to less than 2 m in the shallower sectors. The age of these gravels was determined at the base of an opening over 9 m depth at Liteni (on Siret r.) and was estimated at 20.3 YBP (Rădoane et al., 2015), which is typical for the Late Glacial – Holocene transition;

ii) another sedimentary pile formed next, consisting in finer gravel with a sandy matrix 2 to 5 m thick, which presented an increasing tendency along the rivers. In the upper part of the floodplains we identified a differentiation between Suceava and Moldova rivers on one side and Siret river on the other side. In the floodplain of Siret river, the upper reach is dominated by fine materials, stratified sands and thick deposits of loams, especially on the 5–7 m high floodplain terraces. Instead, on Suceava and Moldova tributaries, the upper part of the boreholes reveal a consistent pile of uneven-sized gravels that extended up to 4 m, and which, as described in the following sections of the paper, are linked to the fluvial activity during certain phases of Mid and Late Holocene periods.

Summing up, the shape of the sub-alluvial valleys as well as the thickness and disposition of the sediments along the longitudinal profile of rivers floodplains reflect, at least for the exemplified rivers,

the closeness of the Carpathian source area and the intensity of the tectonic movements. These two main causes are in charge of the irregular morphology of valleys incision (depths alternance of over 20 m with sectors where the fossil bed is close to the surface) and of the thickness of floodplain sediments deposits. Particular to sediments accumulation in the floodplains of the East–Carpathian rivers is the successive alluvial fans formation mechanism along the rivers, which maintains active in the present. The age of the incision formed in the fossil bed of the valleys has been determined as MIS3, MIS2 (for the VIIth order rivers in Strahler system) and MIS4 (for IInd and IIIrd order valleys in Strahler system) (cf Perșoiu et al., 2017).

5. Floodplains geomorphology and correlation with their stratigraphical structure

In order to obtain a more comprehensive Holocene fluvial history of the investigated hydrographic systems, in this section, based on the stratigraphical information from the field, we made a synthesis of certain sedimentation patterns that are representative for the studied geographical area. The surface morphology of the river floodplains alluvial complexes was coupled to the investigations on the natural and artificial openings in floodplain deposits. In each of the three key sites that we described we benefited of more outcrops in the floodplain deposits based that we used to accomplish the composite stratigraphic sequences. An individual and analytical presentation of each site is given below.

5.1 Verești site, Suceava river

This key site is located at about 15 km upstream of the confluence with the collecting river Siret, and it was chosen for its representative profile of the floodplain deposits built along the 35 km of the channel that were investigated. The floodplain is 2.5 km wide and it is bordered by the 10–15 m terrace (on the left side) and by the hillslope of Suceava Plateau (on the right side). The general morphology of the floodplain was mapped in detail and the results are presented on the geomorphological map of the 10 km long reach (Fig. 4A). The channel of Suceava river is sinuous–wandering with alternating

bars made of gravels with a mean grain size of 15–20 mm. The active belt where the channel migrates without constraints during the maximum annual discharges (cf Bridge and Leader, 1979) represents the lowest level within the floodplain (below 1.5–2

m). Its width in this sector is around 200–300 m. There are several generations of abandoned channels distinguishable over the floodplain surface, which are deepened in three floodplain levels: 2.5–3 m, 3.5–4 m and 5–7 m (Fig. 4A).

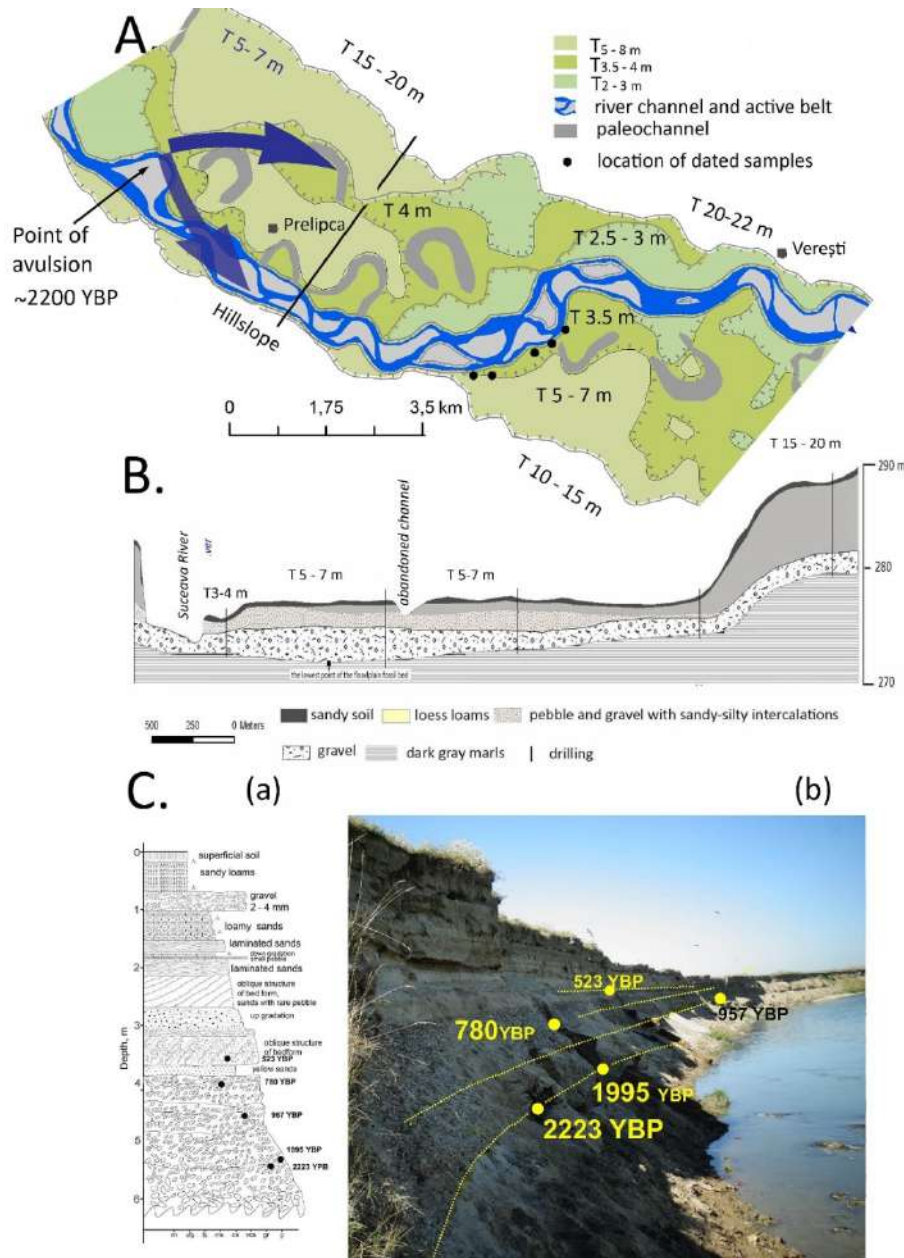


Figure 4 A. Geomorphological map of Verești site reach (to see Fig.1 to locate); B. Cross-section through floodplain fill (modified after D. Oprea, 2014); C. Lithostratigraphical profile (a) and radial gravel lobes in the right concave bank of the river and their ages in YBP (b). See discussions in the text

A series of 5 boreholes performed by SGA (Sistemul de Gospodărire a Apelor, in ro) Suceava (cf Oprea, 2014) on a transverse profile in this sector

(Fig. 4B) allowed us to get a closer look on the sedimentary structure of the deposits. The fossil bed of the floodplain is situated at a maximum depth of

ca. 2 m lower than the present river bed and it is positioned around 1.5 km left of the present course. This fossil bed is formed in the Lower Sarmatian sedimentary complex and it is built mainly of compact purple marls. The coarse gravels that form the basis of the floodplain alluvial deposit are maximum 3 m thick. Over this gravels level there is a 1 to 2.5 m thick horizon of small pebbles with sandy matrix and at the surface there are sands and loessoid loams that lay on the 5–7 m floodplain level. This area was the highest from the mapped floodplain sector, therefore it had the conditions to develop and presently maintain in the Prelipca village territory. Westward of this floodplain level there are surface traces of a former course of the channel, on which we will return.

The loops of the river channel sinuities contain openings in the floodplain deposits that reveal its structure on 5–7 m thickness, down to the bedrock. The profile presented in Fig. 4C revealed a sediment pile comprised between 3 and 6.5 m, finally reaching the bedrock. This makes us believe that the present course of the channel is deviated laterally of the older one. Due to the relatively recent incision of the channel (–1.5 m since 1963 until the present, cf. Rădoane et al., 2013) and to the lateral erosion in the concave banks, the structure of the floodplain deposits was visible and we were able to describe it. The main sedimentary units are visible in the lithostratigraphic profile analysed on the right concave bank of the river (Fig. 4C): a 2.5 m thick unit formed by coarser gravels ($D_{50} = 15\text{--}20\text{ mm}$) at the base (the contact with bedrock) which are getting smaller ($D_{50} = 5\text{--}7\text{ mm}$) towards the upper part; a transition unit 1.2 m thick formed by small gravels (D_{50} smaller than 5 mm) with oblique and laminary structures trapped in coarse sands horizons; a unit of finer sediments, sandy and clayey, with small insertions of small pebbles. Age datings for the sediment deposits from this profile prove that the alluvial processes on a new channel course started 2200 years BP ago. It was then when the river moved its course from the left side of the floodplain (highlighted on the map at west of Prelipca) towards the right slope through an ample avulsion phenomenon. In the following 1500 years, sedimentation continued on the present course of the channel;

gravel horizons deposited interruptedly which buried no less than 15 oak trunks inventoried on this approximately 800 m long reach (Fig. 4A, C). The disposal and age of the trunks demonstrate that the alluvial processes occurred as flood waves that left behind radially disposed gravel lobes (Fig. 4C). After 523 YBP, the sedimentation manner continued, the lobes of small gravels and coarse sands with oblique structure being sedimented on the right bank of the river.

5.2 The Tupilați site, Moldova river

This site is represented by a 10 km long floodplain sector, where the surface morphology develops as follows: the fluvial migration corridor of the channel is over 1 km wide, comprising multiple islands and secondary branches of the channels. In the last 100 years the width of this active belt of the floodplain showed a gradual reduction (with up to 49%, Chiriloei, 2012) as the river channel modifies its typology from braided to wandering. The active belt is bordered by three levels, of 2.5–3 m, 4–5 m and 5.5–6 m (Fig. 5A).

The age estimation for the accumulation that forms the floodplain of Moldova river was possible by combining the numerous radiocarbon datings made on the profiles opened in the gravel pits from this sector. The oldest gravel beds that have a direct contact with the base rock (Sarmatian marls) are 10 ka cal BP old (Fig. 5B) and represent the 4.5–5 m floodplain level that maintains on the left bank, at the contact with T 10–12 m. Within this accumulation an incision of the river occurred, separating the 3.4–4 m level. The lateral migration of the river between 10 ka and 3–3.7 ka cal BP contributed both to the incision of the river and to the redistribution of the sedimentary material from the higher terrace. One of the high amplitude phenomena registered on Moldova river was the generalized alluvial sedimentation along the valley (at least upstream of Gura Humorului) with coarse materials that were deposited as a gravel sheet with an estimated thickness of 3–4 m, which is well individualized also on the resistivity profile (Fig. 5C) in the 3.5–4 m floodplain level. The age determinations based on the oak trunks embedded in this sediments deposit have revealed that the process occurred after epi-

sodes of large floods (Chiriloaei et al., 2012; Rădoane et al., 2015). Besides, the composite lithostratigraphic profile (Fig. 5B) registers this layout of the gravel sheets with a rather chaotic structure, with small insertions of fine materials. On the top of the 3.5–4 m terrace, we identified several abandoned branches of Moldova river filled with

fine sediments deposited in thin layers, their position revealing the high capacity of lateral migration of the river channel. Small floodplain levels of 1.5–2 m were subsequently shaped in the alluvial deposit of the 3–4 m terrace and are now connected to the present fluvial corridor. We estimated their age at 1.2–2.2 ka cal BP.

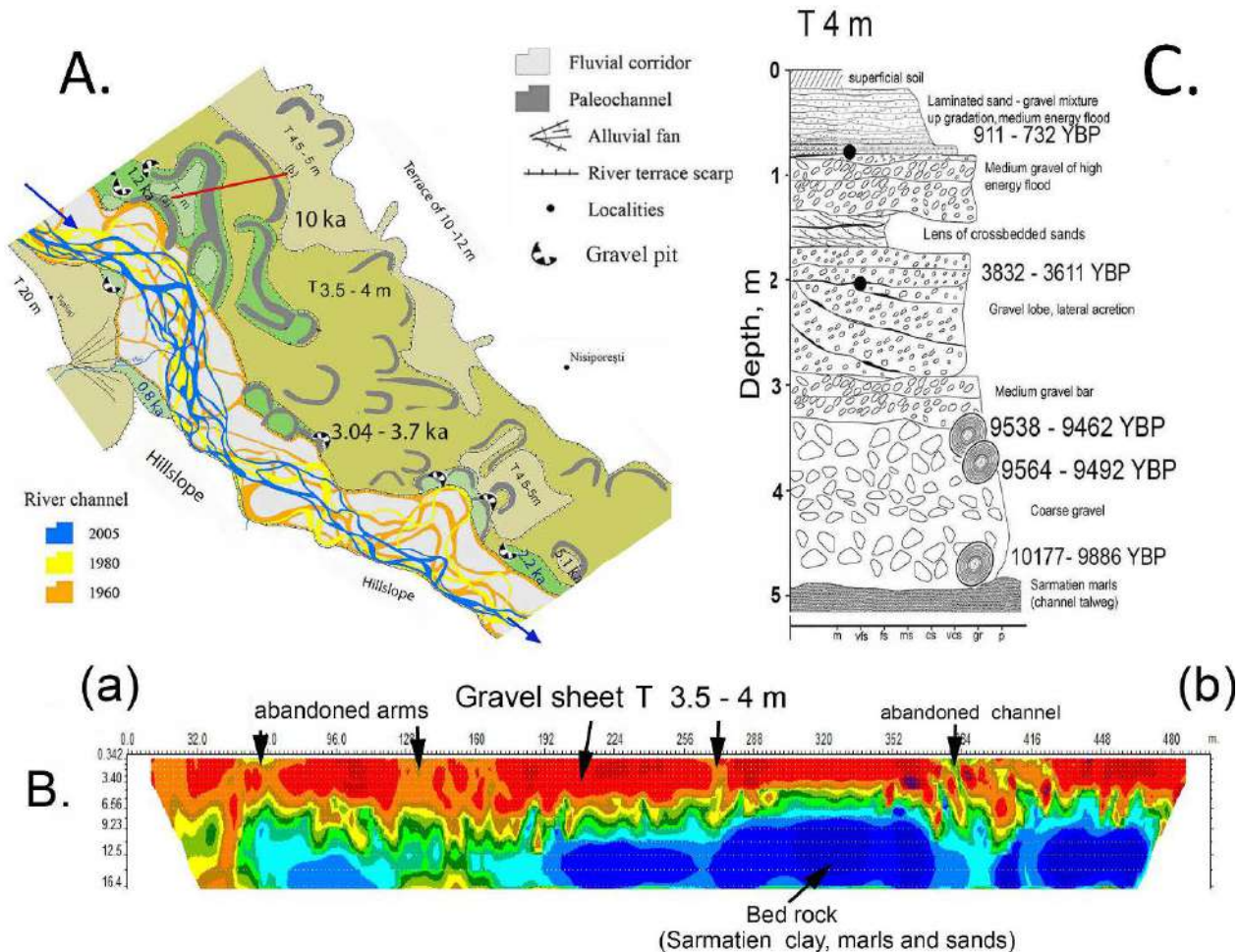


Figure 5 The key site Tupilați, on Moldova river (modified after Rădoane et al., submitted). A. The geomorphological map of the present dynamics of the river channel, the extension of the floodplain terraces and the estimation of their age based on the datings of the sediments deposits from the profiles uncovered in the gravel pits. B. Resistivity profile in floodplain deposits and old courses of the river channel. C. Lithostratigraphic profile representative for this sector

5.3 Liteni site, Siret river

Liteni site is situated just downstream of the confluence of Suceava and Siret rivers (Fig. 6). The site is exemplified by a composite profile which was described based on a 4 km long and 3–7 m thick opening in the floodplain sediments of this river (Fig. 6A) and on a geomorphological map of the floodplain covering ca. 8 km of the valley length

(Fig. 6B). The investigated river sector is characterized by a meandering to sinuous channel, with gravel bed and banks with mixed structure, with lateral accumulation sectors and well developed alternating bars. The surface morphology of the floodplain in this sector is composed of the active belt of a 700–800 m length and three levels of terraces (2–2.5 m, 3–4 m and 5–7 m) with nonuniform distribution. The relief from the top of the terraces

preserves the traces of some old channel loops, presently silted.

The sedimentation history of the river begins with a layer of coarse, slightly cemented gravels which were dated at 20.3 ka cal BP. The level of the gravels ends with an erosion line at a depth of 7.5 m. The contact with the upper finer sediments is marked by a bed of oak logs, the oldest ones reaching 9.5 ka cal BP. The upper sediment deposit is finer and finer towards the surface, and the ages succeed from 6.7 ka cal BP to 2 ka cal BP on the way to the upper part of the profile. The sedimentary unit situated 6–7.5 m deep is formed of lobes with a cross-bedded structure of fine gravel with coarse sands matrix. Their age ranges from 5.2 ka cal BP to 4.9 ka cal BP. These gravels are followed by mean-size yellow sands horizons with rare grains of gravel, the age of which was estimated at about 3.65 ka cal

BP at the upper part of the deposit. The succession of the horizons was interrupted by the incision of a channel, the contour of which (64 m width and reaching a depth of 6 m) could be determined clearly in the lithostratigraphic profile of the 5–7 m level terrace. The paleochannel was filled with fine sediments with laminary structure, rich in organic materials. The age of these sediments in the base of the paleochannel was dated at 2–2.2 ka cal BP. Therefore, we can assume that the paleochannels identified on the surface of the 5–7 m terrace were functional before 2.2 ka cal BP. Afterwards, the channel migrated to the opposite bank, separating the succession of levels with lower altitudes. In the same terrace, but in a different site, we identified a fossil soil profile at 2 m depth below the sandy loams from the surface, which indicates that the terrace was not flooded after 2.2 ka cal BP.

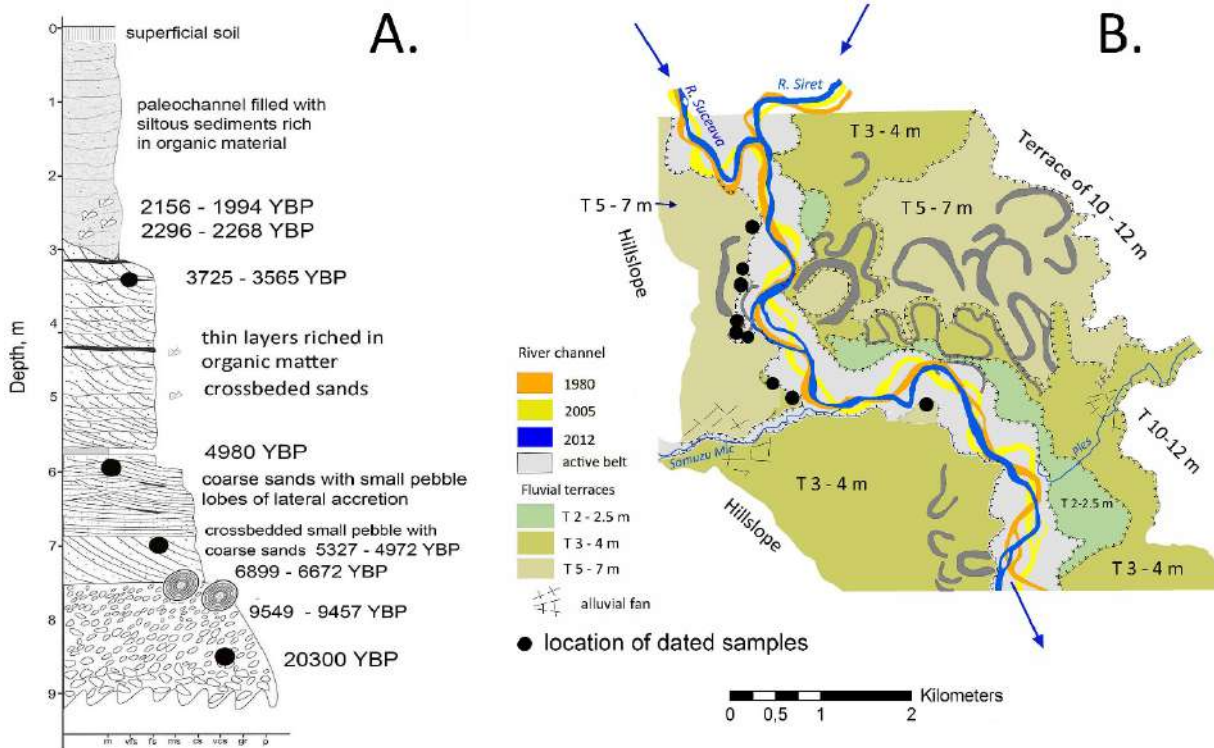


Figure 6 Liteni site: (A) Composite lithostratigraphic profile in the 5–7 m terrace of Siret river; (B) Geomorphologic map of the site and the location of the dated samples. Discussions in the text

6. Discussions and conclusions

The history of sedimentation in the floodplains that we exemplified here starts with coarse gravels deposition which, based on their age (older than 20 ka cal BP in the case of Siret river) were put in place

during the Late Glacial. This level of gravels and boulders proves the presence of high-energy braided water courses, a feature that has been proved and sustained for the evolutionary models of the rivers in our geographical area (Howard et al.,

2004; Perşoiu et al., 2017). On other exemplified rivers (e.g. Moldova), the age of the gravel located at the base of the profiles, where it reaches the bedrock, has been dated at 10.1–9.9 ka and presents a depositional model composed by sediment horizons with maximum grain size of 10 up to 15 cm which decreases gradually towards the upper part (diameters of 1–5 cm). In the inventoried openings we only found the upper part of this level, the present level of the rivers oftenly reaching the bedrock (especially in Moldova and Suceava, process accentuated by the intense exploitation of the bed material).

The contact between coarse gravels (LG) and the transitional gravels (LG – H) as well as other phases of fluvial activity during the Holocene was marked, at almost all the sections that we studied, by a quasi-continuous series of subfossil trunks. The position of these subfossil trunks and the coarse texture of the sedimentary material in which they are embedded, indicate the peak of a major flooding, the covering of the trunks occurring during the decreasing phase of the flood wave.

The sediments belonging to the Holocene are generally finer (coarse sands with crossbedding structures), intruded by horizons of small gravels and organic materials. The rivers closer to the Carpathian source–area (such as Suceava and Moldova) present much coarser Holocene sediments, with a sedimentary structure composed of superimposed gravel bars, lateral accretion lobes and up-fining sedimentation. At Siret river, which flows at a greater distance of the Carpathian source, the fine part of the Holocene deposits is the prevailing one and in the 7 m terrace a fossil soil profile was mapped, which is not very common in the floodplains of gravel-bed rivers.

The number of datings smaller than three per millennium did not allow us to better describe the individualization of more numerous phases of fluvial activity in the Early Holocene, up to 7000 YBP. Past climate reconstructions for the Romanian territory (A Perşoiu, 2017) have shown that the Early and Mid–Holocene were dominated by warm and relatively dry intervals that extended until the 4700 YBP moment, when a sudden change occurred in the climatic regime. After the 4700 YBP moment, the

cold and wet periods prevailed. Consequently, the flooding episodes had a greater intensity during the second half of the Holocene. This is confirmed in this case by the distribution on 100 years frequency classes of the age datings on fluvial units (Fig. 7D). The available age datings of the fluvial units deposited in special hydrodynamic conditions are grouped, during the last 6000 years, in a series of clusters. This finding leads us to the assumption that, the highest probability for flooding phenomena to occur was in the Mid to Late Holocene. Moreover, a second processing technique of the age datings (using Oxcal calibration curves, Reimer et al., 2013) produced the probability function density curves (Fig. 7E) which have the advantage that, by summing and normalization (cf Hoffman et al., 2008), can identify more detailed the high clustering phases of the exceptional hydrological events occurring on rivers at a centennial and multicentennial timescale. Episodes of flooding were identified where the relative probability exceeded the mean and where at least three dates occurred within two centuries in the corresponding part of the date frequency plot (cf Macklin et al., 2010). Therefore, the results presented in Fig. 7E are more relevant comparative to the frequency distributions from Fig. 7D. The peaks that surpass the mean value of the relative probability indicate that during that time the fluvial activity on the rivers had a greater intensity than in other time intervals. These years are indicated on the figure next to the peaks, respectively: 6200, 5300, 4100–4080, 3600, 2300–2200, 1900, 1300, 950, 750, 520 YBP.

The fluvial units on the rivers from Siret basin which were exemplified here, supplemented by other documented reports for Romania (Howard et al., 2004; Perşoiu, 2010; Bonsall et al., 2015; Perşoiu and Rădoane, 2017) lead to the conclusion that periods of alluvial sedimentation correlate well with episodes of climatic cooling, higher rainfall and enhanced river activity.

In order to argument this observation, we combined several records of the variability of paleoclimates on the territory of Romania, and we selected those that are continuous over the entire Holocene, or at least in the Mid- to Late Holocene.

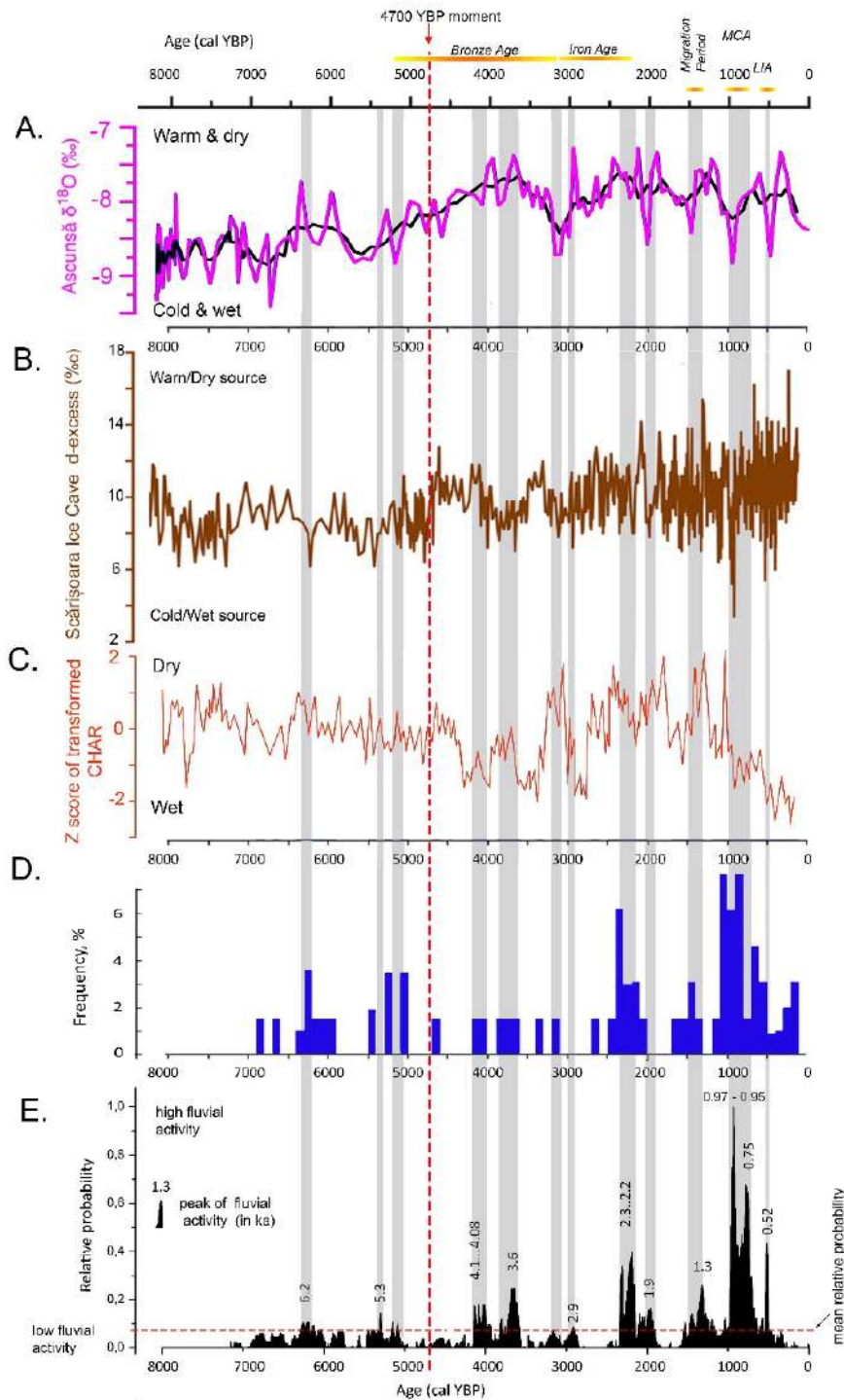


Figure 7 Fluvial activity for East Carpathian rivers and correlations with other paleoenvironmental indicators from Romanian territory in the last 8ka: A. Speleothem $\delta^{18}\text{O}$ records from Ascunsă Cave, Mehedinți, SE Carpathians (Drăgușin et al., 2014); B. Scărișoara Ice Cave d-excess record (Perșoiu et al., 2017); C. Z-score of charcoal accumulation rate at Lake Știucii (Feurdean et al., 2013); D. The frequency of the calibrated ages for the fluvial units of the investigated rivers; E. The distribution of the summed relative probability of the calibrated ages for the fluvial sedimentary units. The years marked on the plot indicate peaks with intense fluvial activity on rivers. Discussions in the text

The proxy data that we used for the paleoclimates analysis were gathered from: speleothem isotop record from Ascunsă Cave (Mehedinți Mts, Southern Carpathians) (Drăgușin et al., 2014); ice stable isotope record from Scărișoara Cave (Bihor Mts, Western Carpathians) (Perșoiu et al., 2017); sedimentary charcoal record from Lake Știucii (Transylvania Plain) (Feurdean et al., 2013). All the three categories of proxy data present a common variability trend, given by the change benchmark from around 4700 YBP (that we highlighted in Fig. 7 through a red line).

A common variability trend is observed at all the three categories of proxy data which show the same change threshold, situated around 4700 YBP (this was highlighted with a broken red line in Fig. 7). A similar tendency is also visible in the case of rivers activity (a greater number of sedimentation events during floods, respectively) in the second half of the Holocene (in terms of both frequency of events, Fig. 7D, and of the relative probability of events, Fig. 7E). The fluvial activity peaks were lengthened with a grey band so that the correspondence with the variability of the other climatic proxies could be followed. It is not our intention to get the perfect adjustment of the wetter and colder phases with the drier and warmer ones. This would be an exaggeration, given the climatic complexity over the territory of Romania and the inherent limitations within the climatic proxy data taken into account. However, we can keep in mind two general observations from the analysis of the diagrams, respectively: (i) different environmental processes (speleothem growth rate, ice accumulation rate in the cave, charcoal accumulation rate providing information about the drivers of fire regimes, fluvial processes intensity) were predominantly controlled by the climatic variability from the Holocene; (ii) during the second part of the Holocene (especially post 4.7 ka) occurred more flooding episodes that coincided with the wet and cold periods identified as well in the variation of the other proxy data (for example, a lower charcoal accumulation rate in the sediments from Lacul Știucii, lower values of the $\delta^{18}\text{O}$ and $d\text{-excess}$ both indicating a cold and humid source of precipitation, smaller values of the

stable isotopes in speleothemes). The climatic explanation of these modifications was discussed by Perșoiu (2017), Perșoiu et al., (2017) and consists in a shift from NAO– to NAO+ in the North Atlantic, the latter becoming dominant since ~4.7 kcal BP (with interruptions most notably, between 4.2 and 2.5 kcal BP).

Returning to the study cases from this paper, we can use these palaeoclimatic information to explain a great part of the major moments of river channels dynamics change. For example, on Suceava river (Fig. 4), the discharges on the valley around 2200 YBP were so high that determined a wide avulsion process, causing the river to change its course direction on a distance of about 3 km and separating, in the middle of the floodplain, the 5–7 m terrace on which Prelipca village is settled today. The traces of the channel previous to 2200 YBP are preserved within the floodplain at the 3.5–4 m level. The new course of the Suceava river channel was kept until present with reduced migration within the fluvial corridor. The most significant gravel depositions with large lobes structure occurred in 950 and 780 YBP and the fine gravels with oblique structure were sedimented in 520 YBP. All these sedimentary units reflect an intense hydrodynamic activity, typical to extremely rainy periods in the river basin, which were documented by the above mentioned palaeoclimatic studies. The moments described above are concurring with the cold and wet climatic phases highlighted by the variation of all the proxy data represented in Fig. 7 and occurred at the end of the Iron Age. The following instability periods reported at Verești site were simultaneous with the so-called Medieval Climatic Anomaly and with LIA.

On Moldova river (Fig. 5), in the described sector, both the quarry pits openings and the resistivity profiles highlighted a wide sedimentation phase with a 3–4 m thick gravel sheet dated with ^{14}C at 3700–3000 YBP, a palaeoclimatic interval favourable to exceptional flooding even in the west of Romania (Perșoiu, 2017). This interval occurs during Bronze Age and the climatic proxies indicate, more or less, a climatic period favourable to flooding (Fig. 7).

On Siret river, the wet phases of the Holocene manifested through floods that determined the

accumulation of lateral accretion lobes (formed by small gravel that succeed from 5.2 ka cal BP to 4.9 ka cal BP, typical to a sinuous to meandering channel as the one of Siret river). The 2200 YBP moment has also been marked in the sedimentary archive of Siret river at Liteni, by the incision of a channel in sediments older than Mid-Holocene which was subsequently filled with finer flood sediments (Fig. 6A). Using both the structure of the dated floodplain sediments and the identified paleo-terraces on floodplain surface we estimated a lateral channel migration rate of 700–1000 m, while the alluvial material was re-worked, especially through lateral accretion processes (Rădoane et al., 2015). The lateral migration process was accompanied and boosted by avulsion phenomena (as we showed on Suceava river), the frequency of which is connected to the periods with exceptional pluvial and hydrological events.

The fluvial sedimentation activity, intensified by the exceptional hydrologic events, has been intercalating with the intervals when rivers deepened in their own sediments, this phenomenon leading to the formation of floodplain terraces. In the case of the rivers that we investigated, three or four generations of floodplain terraces were identified at the 5–7 m, 3–4 m and 2.5–3 m, the active belt on which the present channel oscillates being situated below 2 m height. The exact dating of the incision phases is more difficult to obtain. The estimation of floodplain terraces detachment was made by correlating this process with the warmer and drier intervals of the Holocene which intercalated with the wet ones. The warmer intervals, with a stronger vegetation cover and only slight human activity were optimal for river channels incision. The deepening and sedimentation did not exclude each other along time. On the contrary, during the colder time intervals, along the sediment deposition processes there were also river channels incisions and great lateral mobility (avulsions), whereas in the warmer time intervals, except for a generalized incision, local sedimentation phenomena and reworking of the alluvial sediments were also possible. Our results as well as the ones documented by other studies (Macklin et al., 2006; Starkel et al., 2006; 2012; Perşoiu, 2010; Panin and Matlakhova, 2015; Perşoiu and Perşoiu, 2018),

show that the process of river channels deepening was prevailing during the Early and Mid-Holocene and that the sedimentation, avulsion and incision processes had a much more pronounced activity rhythm in the second part of the Holocene (Rădoane et al., 2018).

The conclusion of these discussions draws the attention upon the interpretation potential given by the fluvial archives in deciphering the Holocene sedimentation history. In this case, we could identify a general pattern of this history for the East-Carpathian rivers, by putting together the sedimentation phases during the Mid- and Late Holocene, the avulsion and lateral migration processes that accompanied the high fluvial activity periods and the patterns of river channels morphological adjustment from braiding to wandering and to sinuous-meandering. The chronological overlay of the events indicates that during the Mid- and Late Holocene, the time intervals with the most ample fluvial processes on the rivers that drain the Eastern Carpathians were (in BP years): 6200, 5300, 4100–4080, 3600, 2300–2200 (Iron Age), 1300 (Migrations Period), 950–970 (Medieval Climate Anomaly), 750 and 520 BP (Little Ice Age) and that these intervals generally overlay the palaeoclimatic episodes characterized by high amounts of precipitation. Coming to more recent years, for example, the last 200 years, the investigations regarding river channels dynamics (especially plane configuration change) have indicated for the first time an unprecedented increase of the anthropic component weight amongst the causes of the phenomenon. Consequently, the human impact is now competing with climate changes for the greatest share in controlling fluvial processes (Perşoiu and Rădoane, 2011; Rădoane et al., 2013, 2017).

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Fuzzy modelling of slope erosion by runoff. Case study in Corinth basin, Greece

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ABSTRACT

This research applies the fuzzy set theory via Geographical Information Systems (GIS) – based analysis, in order to model slope erosion by runoff. For this purpose, the following steps were accomplished: definition of input variables (rocks' susceptibility to erosion, slope inclination, slope morphology), development of a fuzzy inference system based on theoretical and empirical knowledge, transformation of input to output variables (erosion – deposition) and visualization of the output variables (spatial distribution of erosion–deposition processes). The method was applied on the drainage basin of Corinth, located in the north–eastern part of Peloponnese (Greece), where a series of catastrophic erosional events have recently occurred.

KEYWORDS

fuzzy modelling; slope erosion; GIS–based analysis; Greece

1. Introduction

Erosion rates depend on a variety of factors, such as the frequency and magnitude of precipitation, surface and subsurface waters, slope morphology (length and inclination), rock's susceptibility to erosion, land use and human impacts.

The amount and intensity of precipitation control surface and subsurface runoff. Splash erosion by raindrop impact results in the breaking and detachment of soil particles. In addition, slope char-

acteristics have a significant role. Sediment removal is estimated as a function of distance to divide multiplied by a power of slope gradient (Kirkby, 1978). Rock composition and texture, expressed as physical resistance, can further affect erosion rates (Meybeck, 1987).

The vegetative cover acts as a protective mechanism to the weathering material by reducing the energy of rain, but also controls the soil and the

weathering mantle thickness and, consequently infiltration.

The increasing human interventions to the environment have contributed to the acceleration of erosion processes, for instance with frequent forest fires and urban expansion. Human interventions also control the vegetative cover, through the management of cultivated areas and pasture lands (Evelpidou et al., 2018). Moreover, changes in land use, such as expanding agricultural land and deforestation, have made the weathering products more vulnerable to erosion. In terms of runoff erosion, three factors of land use are critical, the vegetative cover and its role, the surface cap and the mechanical disturbance of the soil (Wischmeier, 1975; Renard et al., 1997; Weltz et al., 1998).

Another important variable that needs to be taken into account is climate change and its causes. The International Panel on Climate Change (IPCC) has predicted frequent and severe droughts and floods caused by irregular climatic conditions in the future, making water resources management difficult (Ahn et al., 2018).

Runoff may be expressed as sheets of water or as small rills; its eroding capacity being related to the vegetative cover and it is expected to gradually increase downslope (Langbein and Schumm, 1958). During heavy precipitations, gullies tend to grow in depth and upslope. This process leads to the expansion of the drainage basin. The transported material can be deposited in the low catchment areas or may end up in a stream channel. This highlights the close interrelations of stream and slope evolution.

Weathering and erosion processes are also closely interrelated. The weathering mechanism produces new material, while erosional processes transport the material to the lower parts of the drainage basin. Erosion is active, as long as surface deposits are present, and when the whole waste mantle is removed, erosion agents become inactive. This mass balance (Ahnert, 1987) controls the spatial distribution of the denudation weathering-limited and denudation transport-limited areas in the drainage basins (Young, 1972; Chorley et al., 1984). It is noted that in slope hollows, the material is slowly transported downwards by creep processes, while rapid mass movements act as coupling points

by removing the material (Ahnert, 1996; Selby and Hodder, 2000). According to Wolka et al. (2015), the combination of steep slope, low vegetation cover, along with erosive rainfall or extreme weather conditions, induces a non-tolerable soil erosion problem. Therefore, erosion risk mitigation strategies should focus on the protection of steep slope areas.

In this context, the aim of this paper is to study the dynamic processes of erosion by mapping zones of different erosion degree and deposition in a holistic approach. This mapping is related to practical problems caused by nowadays erosion events, and its scope is formulating measures to stop or limit flow concentration and minimize erosional events (Kirkby, 1978). Nowadays several models have been developed for mapping erosion, however there is a lack of precise methods of monitoring. Traditional field-based methods of monitoring erosion (sediment traps, erosion pins, and bridges) can be labour intensive and therefore are generally limited in spatial intensity and/or extent (Gillan et al., 2016). Therefore, this work also aims to stimulate the necessity of developing a long-term monitoring mechanism for the repeated floods and soil erosion events in order to protect environmental deterioration.

2. Regional setting

The study area, the drainage basin of Corinth, is located in the north-eastern part of Peloponnese (Fig. 1). This area was chosen as a case study mainly because it is dominated by vulnerable rocks prone to erosion, and, because catastrophic floods accompanied by vast erosion occurred recently (1997) and are expected to happen again.

The drainage basin of Corinth is characterized by a low-lying coastal area in the northern part and a hilly and semi-mountainous area in the south. The area is mainly composed of Quaternary and Neogene formations at the northern part and limestones of Alpine age at the south. The Neogene formations consist mainly of alterations of marls, sandstones, conglomerates, limestones, etc. (Bor novas et al., 1972). The drainage system of Corinth is characterized by an elongated basin, of a north-south direction, with a dendritic form. The form and

size of the drainage system are related to the topographic, geological and hydrological parameters.

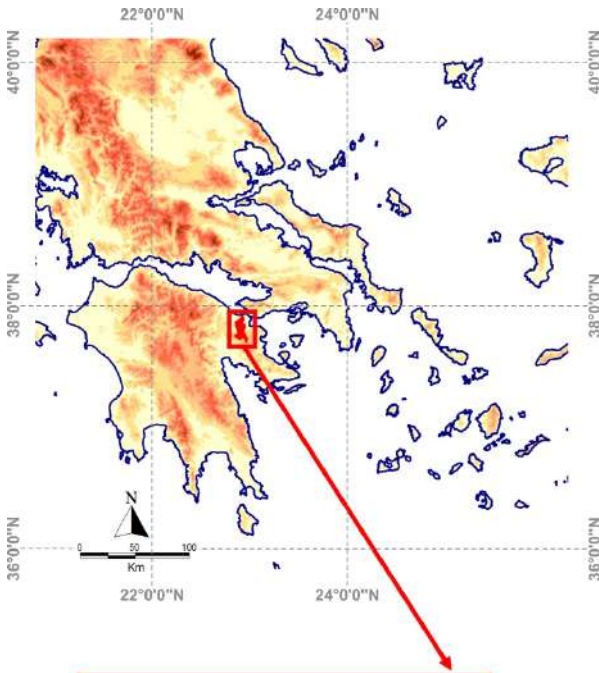


Figure 1 Location of the study area

3. Methodology

The study of erosional processes demands a series of steps in order to obtain the final output maps. These steps include (Fig. 2): 1) field-work, interpretation of aerial photos and satellite images, development of a GIS database with geological, topographical and drainage data), 2) definition of input and output variables, 3) formulation of appropriate logical rules (boolean or fuzzy) for the transformation of input to output variables, 4) analysis and visualization of the output results.

For this work, the input variables used were susceptibility of rocks to erosion, slope inclination and slope morphology. The output was the erosion – deposition variable. In previous research, Gournelos et al. (2002, 2004) have proposed an empirical way to estimate the rock's susceptibility to erosion. The susceptibility of rocks depends on a number of variables, with spatial and temporal variation. Firstly, it depends on the physical and chemical characteristics of the rocks. Thus, mineral composition is the determining factor for chemical weathering. Olivine, augite, hornblende, biotite and dark-colored minerals in general are more susceptible to weathering processes than orthoclase, muscovite, quartz and other light-coloured minerals (Sparks, 1965). Secondly, it depends on the presence of a protective vegetative cover, which contributes to the increase of the infiltration capacity and limits runoff. In general, the grain's form and size define permeability of rocks and, consequently, they control the time of water saturation and the initiation of runoff. Three categories of permeability values may be distinguished: very low (10⁻¹² – 10⁻⁸ m/s), low to medium (10⁻⁸ – 10⁻⁵ m/s) and high (10⁻⁵ – 10⁻² m/s) (Bolton, 1979). Thirdly, it depends on the characteristics of the involved processes: the intensity and duration of precipitation, its spatial and temporal distribution and the raindrop size. Susceptibility of rocks to erosion is difficult to be directly measured. There are some observations concerning either rock resistance to abrasion (Kuenen, 1956) or mass strength classification (Selby, 1987). The first attempt to assign susceptibility values to various rock types was made by Jensen and Painter (1974).

The second processed variable was slope gradient—there is a linear or power relation between slope and inclination. When slopes are steeper, both the gradient and erosional products (Hack and Goodlett, 1960; Schumm, 1977; Kirkby, 1969, 1978, 1995). gravitational movement and downslope sediment delivery are higher. It has been noted that, in general,

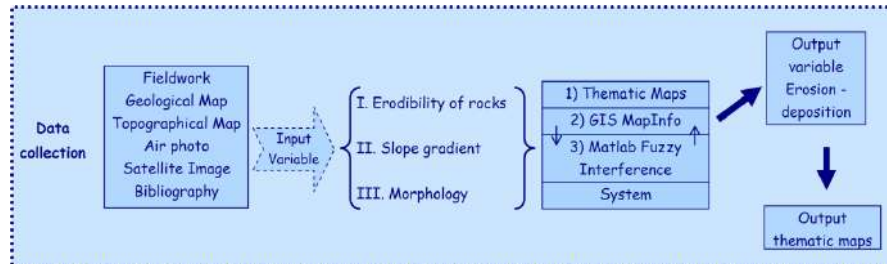


Figure 2 Flow diagram of the methodology followed in this work

The last input variable, slope morphology, is used to reflect the slope form (convex, concave, straight), the slope surface roughness and vegetative cover, the relation (transition) of a certain slope segment to upslope and downslope elements; in total, it defines the degree of the dominant erosional or depositional processes.

A large database was developed based on data derived from fieldwork and existing maps. In order to process different data sets, all input data were

normalized. The entire original dataset was normalized by dividing them by their maximum value, in order to facilitate the comparison of data layers. The input variables (rocks' susceptibility to erosion, slope gradient and morphology) were treated as fuzzy ones (Zadeh, 1965, 1987; Yager et al., 1987; Dubois and Prade, 1980; Zimmermann, 1991; Klir and Yuan, 1995) assigning them the proper membership functions (Fig. 3).

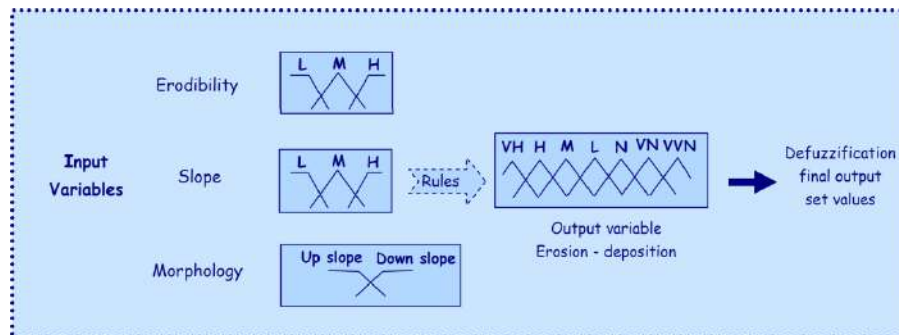


Figure 3 Flow diagram of the form of the membership functions of the input and output variables

The next step was to construct a fuzzy inference system transforming input to output variable (erosion–deposition variable). This was achieved by the formulation of the proper conditional rules based on theoretical as well as empirical knowledge of the erosional processes (Table 1).

The implementation of all the aforementioned procedures was achieved, firstly, by using the GIS MapInfo platform for the analysis and visualization of the input and output variables and, secondly, by using the MATLAB software package to construct the fuzzy inference mechanism based on Mamdani methodology (Mamdani and Assilian, 1975). The

output variable, erosion, expresses the degree of erosion and deposition on the drainage basin. This variable is also of a fuzzy nature. The output variables have been subjected to centroid defuzzification mechanism in order to obtain the final non-fuzzy set of values. The first variable, rock's susceptibility to erosion, is closely related to the lithology. Limestones are characterized by high permeability values and are the oldest and hardest rocks in the drainage basin. Their relief is relatively higher in comparison to the northern Neogene and Quaternary formations.

Table 1 The fuzzy logical rules transforming input variables to the output one

If Erodibility Is Low or Slope Is Low	&	Morphological Factor	Is Upslope	→	Erosion Risk	Is Low
If Erodibility Is Medium & Slope Is Medium	&	Morphological Factor	Is Upslope	→	Erosion Risk	Is Medium
If Erodibility Is High & Slope Is Medium or Vulnerability Is Medium & Slope Is High	&	Morphological Factor	Is Upslope	→	Erosion Risk	Is High
If Erodibility Is High & Slope Is High	&	Morphological Factor	Is Upslope	→	Erosion Risk	Is Very High
If Erodibility Is High & Slope Is High	&	Morphological Factor	Is Downslope	→	Erosion Risk	Is Negative
If Erodibility Is Low or Slope Is Low	&	Morphological Factor	Is Downslope	→	Erosion Risk	Is Very Negative
If Erodibility Is High & Slope Is Medium or Vulnerability Is Medium & Slope Is High	&	Morphological Factor	Is Downslope	→	Erosion Risk	Is Very Negative
If Erodibility Is Medium & Slope Is Medium	&	Morphological Factor	Is Downslope	→	Erosion Risk	Is Very–Very Negative



Figure 4 Spatial distribution of rocks' susceptibility to erosion variable of Corinth's drainage basin. The numbers in the brackets represent the frequency of each value range (number of grid cells belonging to the value range)

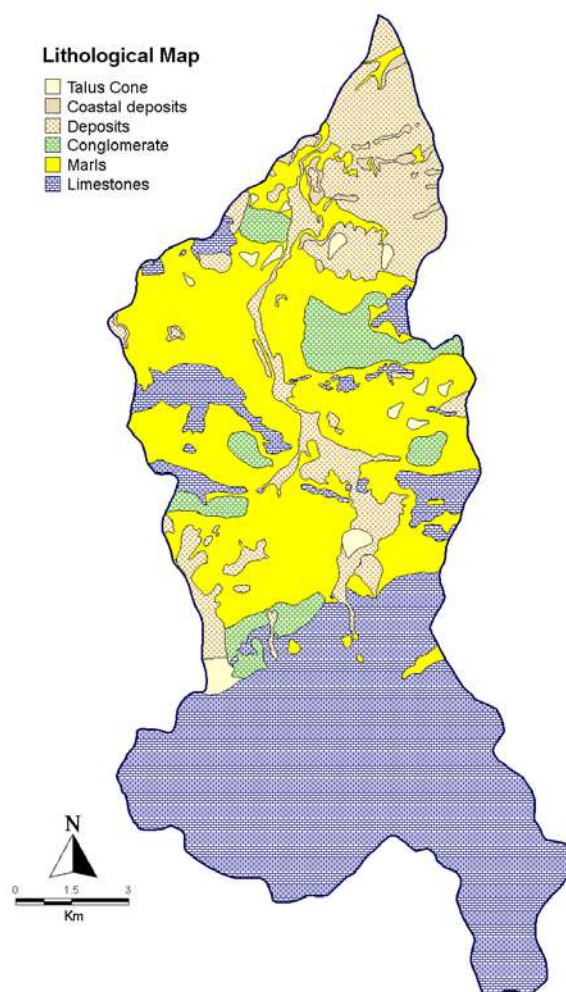


Figure 5 Lithological map of the drainage basin of Corinth

Their relief is relatively higher in comparison to the northern Neogene and Quaternary formations. However, even in the latter group of sediments there are beds of different resistance to erosion and also lithological contrasts within the same bed formations. Conglomerates and sandstones are affected by chemical weathering, when cemented with calcium carbonate or iron oxides. Clays and shales are very susceptible to physical weathering due to the presence of primary or secondary discontinuities and their ability to absorb water (Sparks, 1965).

Three fuzzy classes were assigned to the rocks' susceptibility to erosion for the Corinth drainage

basin, corresponding to the linguistic terms of 'low' (0, 0.5), 'medium' (0.25, 0.75) and 'high' (0.5, 1) susceptibility (Fig. 4). These represent the alpine limestones, resistant beds of Neogene formations and marls, sandstones and Quaternary sediments, respectively (Fig. 5).

The second processed variable was slope gradient (slope form and the predominant erosion or deposition processes), which has been calculated from the digital elevation model. In a similar way, three fuzzy classes have been assigned to slope variable: low, medium and high (Fig. 6).

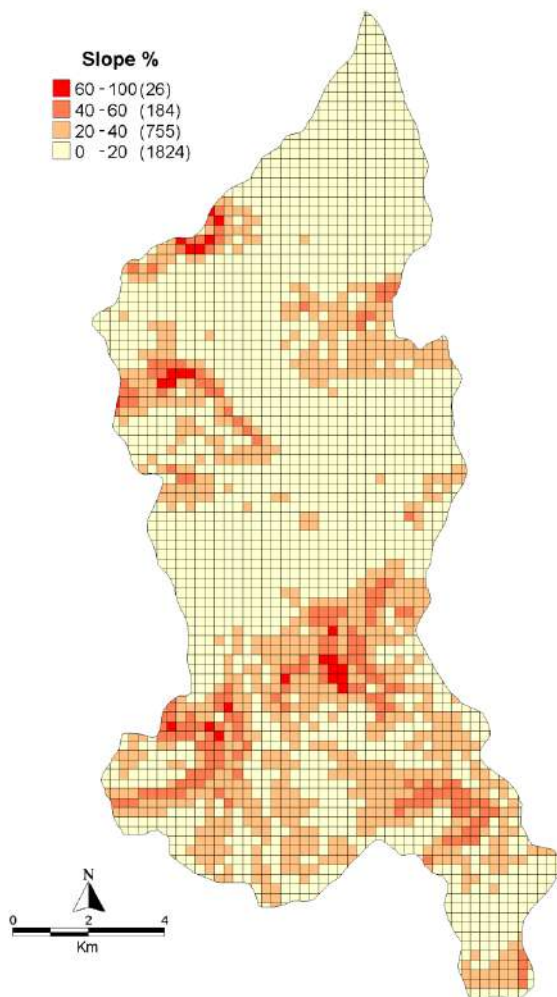


Figure 6 Spatial distribution of the slope gradient variable of Corinth drainage basin. The numbers in the brackets represent the frequency of each value ranges (number of grid cells belonging to the value range)

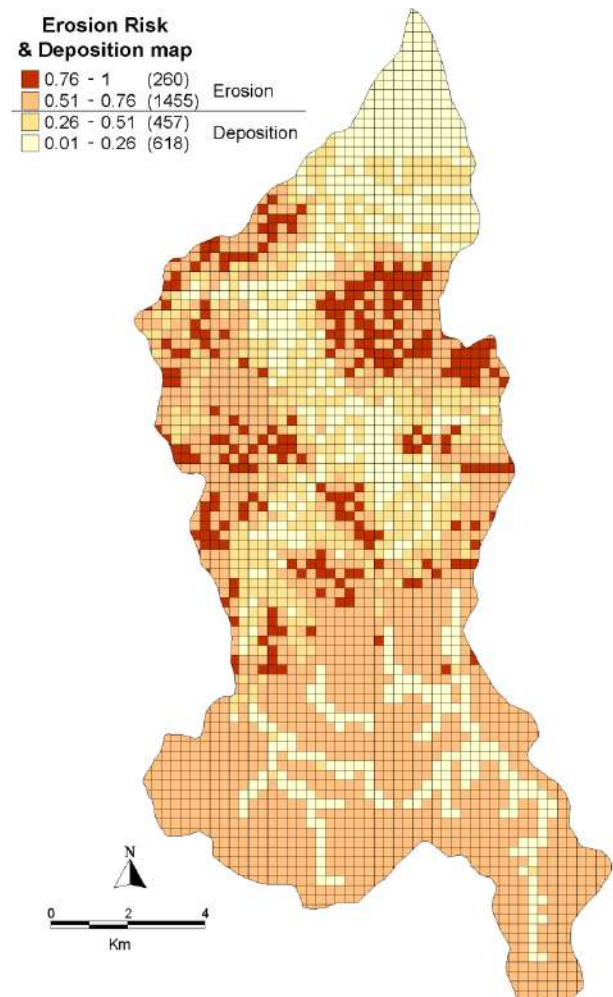


Figure 7 Spatial distribution of erosion – deposition of Corinth drainage basin. The erosion and deposition terms in the legend includes all the grades of this variable as defined in the text and figures. The numbers in the brackets correspond to the frequency of each value range (number of grid cells belonging to the value range)

Slope morphology was the third variable. Apart from the slope gradient, the convex or concave form of the slope, the aspect and the total length of the slope are also very important factors. This variable has two fuzzy classes of values: the upslope and the downslope. Thus, values of 0.6–1 represent convex slope forms, where deposition-limited processes are mainly present, while values of 0–0.6 signify erosion – limited processes domination. The output variables were the erosion–deposition of fuzzy values and take the linguistic terms: Very High, High, Medium, Low, Negative, Very Negative, and Very–Very Negative (Table 1).

Finally, a sensitivity analysis was accomplished on the proposed model, through minimal modifications of the input variables, and the results (outputs) of this process have shown no significant changes to the expected values.

4. Discussion

The main disadvantage of erosion risk maps is that they reflect a static situation, which may change over time; even more importantly is the fact that the deposition-dominated areas are not mapped. Precipitation of variable intensity and duration may produce different erosional events and each erosional event is composed of erosional and depositional zones.

Climate change and extreme weather events call for the development of relatively simple methods to identify various natural hazards, globally; there is therefore need to predict areas of extensive soil erosion using real-world, input data at low cost with considerable accuracy (e.g. Mitra et al., 1998). A holistic approach was applied in this work, by mapping zones of different degree of erosion and deposition. This mapping is related to practical problems caused by contemporary erosional events and it may be used to formulate a series of suitable management solutions in order to avoid or diminish flow concentration and minimize erosional events (Kirkby, 1978).

It should be noted that recent erosion studies have been undertaken in the Mediterranean area (Onori et al., 2006; Ozcan et al., 2008) and new techniques have been proposed to evaluate erosion rates (Wang et al., 2008).

In the case of the drainage basin of Corinth, the resulting map (output variables) is shown in Fig. 7. This is a fuzzy-based approach of the main erosion and deposition zones of the study area.

It is clear that erosion occurs in upslope areas where Neogene or Quaternary formations are mainly present, while deposition occurs in the low flat areas close to the foot slope and the main drainage system. The aforementioned results have been further confirmed through field observations in the study area. This spatial distribution of erosion and deposition zones explains the recent catastrophic flooding of Corinth city, situated in the northern part of the drainage basin. It must be noted that recently, on 11th February 2007, this drainage basin has suffered another flooding event followed by widespread erosion, confirming our estimation about the response of this system to high precipitation.

5. Conclusions

This paper focused on a fuzzy based methodology through GIS based analysis for studying runoff erosion risk, which obliterates the main disadvantage of erosion risk maps reflecting a static situation. The methodology was applied in Corinth drainage basin, and our results were confirmed by field data.

The resulting map of the proposed methodology is a fuzzy-based approach of the main erosional and depositional zones of the drainage basin of Corinth. This methodology is relatively simple to implement using a GIS platform and it could be used together with a long-term monitoring mechanism for the repeated floods and soil erosion events in order to prevent from environmental issues as well as economic losses.

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Mid and Late Holocene evolution of Brateş Lake region (Danube floodplain) based on the multiproxy analysis

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ABSTRACT

This study proposes a local paleo-landscape reconstruction of the Danube floodplain based on a stratigraphic sequence retrieved from Brateş Lake which, by its emplacement near the confluence of Danube – Prut rivers, was fully receptive to changes associated to hydrological, geomorphological or anthropogenic driven events. Due to its intermediate position within the Lower Danube valley Brateş Lake is a proxy for the evolution of Cotul Dunării area (the region of Danube valley turning from S–N to W–E direction) and provide valuable information about the timing of Danube river advancement to the Black Sea after its reconnection to World Ocean. The sediments were analysed to get the history of their deposition by means of accelerator mass spectrometry (AMS) ¹⁴C dating, grain-size parameters, organic matter and carbonate content, magnetic susceptibility together with paleo-fauna and pollen content which altogether led to the identification of main stages: i) delta front advance into Danube estuary (before 8000 BP), ii) shoreline foreshore deposits which describe shoreline position (8000–7900 yrs BP), iii) river floodplain development (7900–5300/5000 yrs BP), iv) lake formation (5300/5000 yrs BP – present).

KEYWORDS

landscape changes, aggradation, chronology, stratigraphy units

1. Introduction

The complex information and the highly diverse proxy elements archived within the preserved sediments render fluvial systems as one of the most intensively studied natural systems on the Earth surface. Since the middle of the 20th century, the number of studies looking for the answer of sedimentary fluvial systems to past environment changes grew considerably and were largely focused on three main issues: the influence of tectonic movements (Molnar and England, 1990; Merritts et al., 1994; Burbank and Pinter, 1999; Ruszkiczay-Rudiger et al., 2016; Arzhannikova et al., 2018), imprints of the climate changes and of the sea level fluctuations (Probst, 1989; Pillans et al., 1998; Blum and Törnqvist, 2000) and the complex links with the human activities and settlements (Gebica et al., 2013; Wilkinson et al., 2014; Jotheri et al., 2016).

Most of the studies concern on a single major event, e.g. the Late Glacial – Holocene transition (Starkel et al., 2007; Turner et al., 2013), the fast sedimentation induced by human activities in the last millennium (Kalicki, 2000; Hoffmann et al., 2008; Kaplan et al., 2009; Morin et al., 2011), with an acceleration in the 19th century and the first half of the 20th century, followed by a sudden decrease as a result of dams constructions in the river basins (Owens et al., 1999; Panin and Jipa, 2002; Preoteasa et al., 2016; Stacke et al., 2014). Recently appeared new studies that treat the avulsions history on long periods of time during Holocene (Tigris and Euphrates: Morozova, 2005; Jotheri et al., 2016; Saskatchewan: Smith et al., 1989; Smith et al., 1998; Morozova and Smith, 1999; Mississippi: Aslan and Autin, 1999; Rhin: Törnqvist, 1994; van Dinter et al., 2017; Meuse: Stouthamer and Berendsen, 2001) or their types and morphodynamics: progradational, reoccupation and anthropic avulsions (Törnqvist and van Dijk, 1993; Morozova and Smith, 1999; Morozova, 2005).

In Romania there is a recent soar for the studies based on the chronostratigraphy of the fluvial archives, especially for the inner medium-size rivers (Teleorman: Howard et al., 2004; Someșul Mic and Transylvanian rivers: Perșoiu and Rădoane, 2011; Perșoiu et al., 2017; Siret: Rădoane et al., 2015, 2018, this volume), whereas those focusing on the Danube

are mostly referring to the Danube Delta (Panin, 2003; Giosan et al., 2006, 2012; Vespremeanu-Stroe et al., 2013; Preoteasa et al., 2016, 2018; Vespremeanu-Stroe et al., 2016, 2017a, b) or to the Danube terraces (Armaș et al., 2018). Recently, a German team which developed a geoarchaeological study on the Neolithic site of Pietrele proposed a local and then regional valley paleogeography reconstruction for Pietrele and Giurgiu–Oltenița sectors (Nowacki and Wunderlich, 2012; Benecke et al., 2013; Nowacki et al., 2018).

The present study also aims to explain the combined impact of the sea level rise and human activities on the evolution of Danube Floodplain in the Brateș Lake area after the reconnection of Black Sea with Planetary Ocean as well as its level fluctuations which induced important fluctuations in the local level of the groundwater level. To prove this, a full stratigraphic sequence was retrieved from Brateș Lake which, by its emplacement within the Lower Danube valley and its position near the main fluvial channel (Fig. 1) was fully receptive to changes associated to hydrological, geomorphological or anthropogenic driven events. The sediments were analysed to obtain the general chronologic framework of their deposition by means of ¹⁴C dating, textural characteristics, palaeofaunistic load in order to get the general picture of the Mid and Late Holocene environmental changes.

2. Methodology

This study is based on a 8 m long core (Fig. 2) performed with an Eijkelpamp percussion corer system into the southern part of the Brateș Lake, which is drained today and used as arable land. In the present, Brateș Lake has a surface of ca. 20 km², much smaller than at the beginning of 18th century (Cantemir, 1716), when it had around 100 km². From the same historical source we found that the lake was fed with sediments from Prut during rainy seasons, through a small channel called Pruteț, and from the Danube at large floods (Cantemir, 1716).

The sediment was sampled in 1 m long PVC pipes, each of them being labelled and sealed immediately after their retrieval and refrigerated at 4°C. The pipes have been opened and described in the

laboratory. Subsequently, 175 samples of sediment have been collected at ca. 20 cm equidistance and corresponding to each stratigraphy change for textural, loss on ignition (LOI), geochronology,

magnetic susceptibility and micro-fauna analyses (for example between 425–500 cm were collected 9 samples for grain-size and LOI analyses and 1 sample for dating).

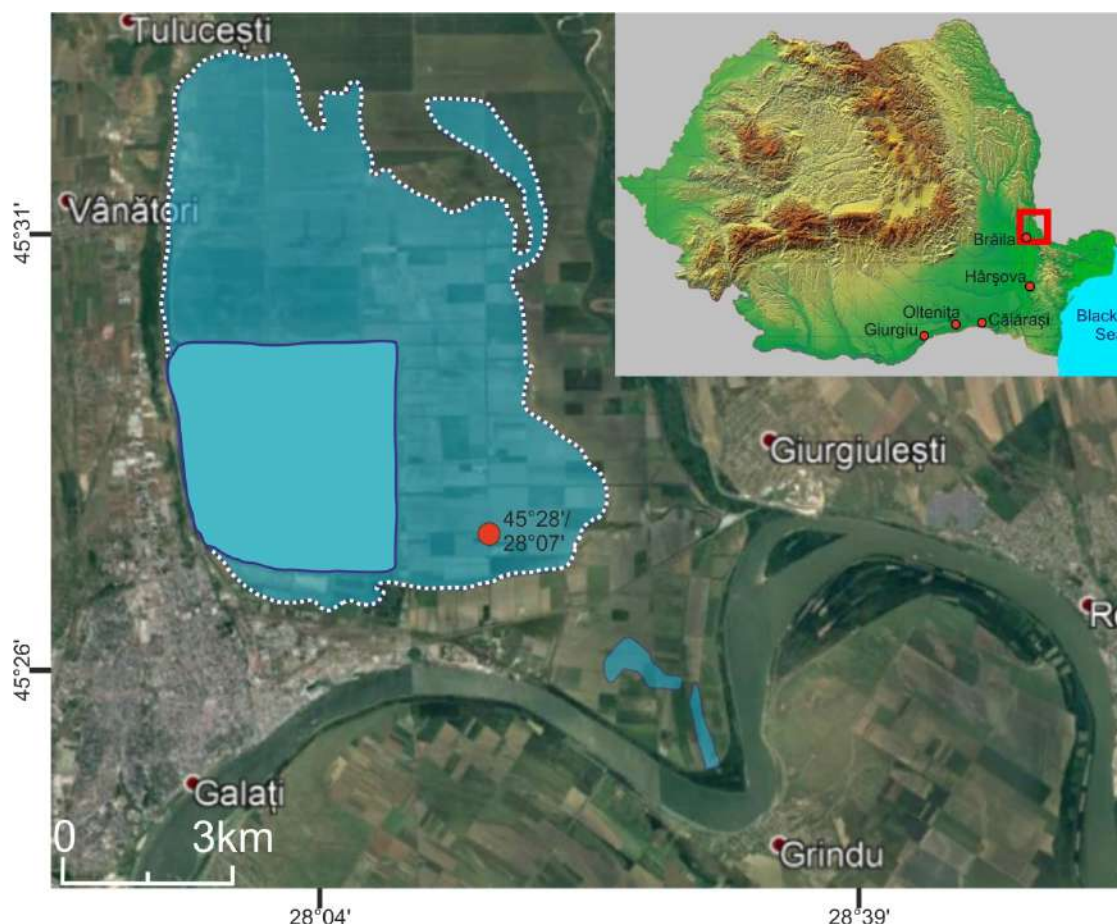


Figure 1 Present location and the former extent (1911; dotted white line) of Brateş Lake before partly desiccation (occurred in 1947–1949), overlapped on 2016 Google Earth image. The red circle marks the position of the core site. Inset: Position of the study site within the Lower Danube Valley / Romania

The chronostratigraphy of the cores was established on five independent AMS radiocarbon dates derived from vegetation fossils/remains, including one peat sample and one small piece of wood (Table 1). Radiocarbon dating measurements were undertaken at two different laboratories: three samples at Gliwice Radiocarbon Laboratory (Poland) and two at Horia Hulubei National Institute for R&D in Physics and Nuclear Engineering (Romania) (Sava et al., 2018). ^{14}C ages were subsequently calibrated using the OxCal online: (Ramsey, 1995), <https://c14.arch.ox.ac.uk/oxcal/OxCal.html>.

Grain-size analysis was realized on 78 samples which, before being measured with Horiba LA950

(Laser Diffraction Particle Size Analyzer), were attacked twice with 20% acetic acid (CH_3COOH), washed twice with distilled water and then attacked for several times with 10% hydrogen peroxide (H_2O_2) to destroy the organic matter content. Then they were washed with distilled water 3 times. Before each reading, several milligrams of sodium polyphosphate were added to each sample to disperse the particles. The calculations on raw data were performed in the GRADISTATv8 program (Blott and Pye, 2001) using the equations developed by Folk and Ward (1957) for grain-size parameters such as mean, sorting, skewness and kurtosis.

Loss on Ignition (LOI) is a cheap and quick method to calculate organic carbon or organic matter and inorganic carbon content. 75 samples, weighing between 2.72 and 4.13 grams, were dried at room temperature for 36 hours. The crucibles were dried at 150°C for 1.5 hours, then cooled and weighed. After that the sediment was added and burnt in three steps: at 105°C for 12 hours (to determine the water content); 550°C for 6 hours and 950°C for 2 hours in a Caloris L1003 oven. After each burning the samples were weighed with an accuracy of 0.0001 grams. To determine the amount of organic and inorganic carbon, the formulas developed by Heiri et al., (2001) were used.

Magnetic susceptibility (MS) was performed at 58 points with a Bartington 3MS device to determine the magnetic susceptibility of the different layers. In general, small susceptibility is associated with an increased content in organic matter (diamagnetic), while high values are associated with the terrigenous sediment flux (ferromagnetic) (Hatfield et al., 2013).

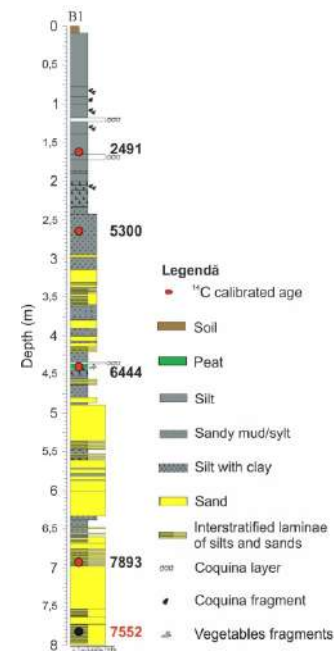


Figure 2 Stratigraphy and chronology of the core based on texture properties and calibrated radiocarbon dates. The deepest sample (Bra_69) was considered an outlier due to the very small size (below the standards)

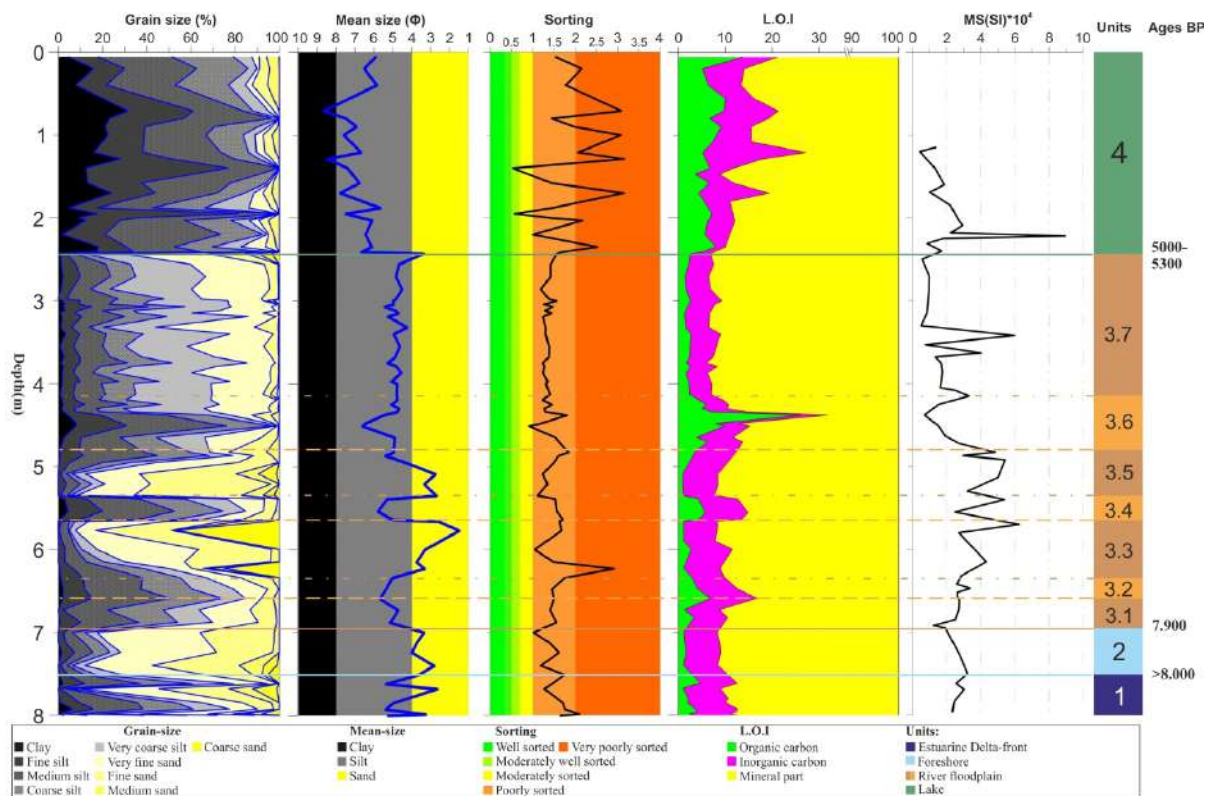


Figure 3 The physical and magnetic proprieties of the Brateş Lake sediment record: percentage of each grains-size class, mean grain-size, sorting, percentage of mineral and LOI (organic matter and carbonate content), mass specific magnetic susceptibility. The right coloured bar expresses the succession of different facies; all the ages are expresses in years BP

Micro-fauna and pollen analyses were done at the Faculty of Geology in Bucharest and at Lakehead University from Canada, on more than ten samples.

Microfaunal analyses were carried out on samples from each sedimentological facies, to infer water paleosalinity and depositional environments. Samples were oven dried at 60°C, weighed and then wet-sieved using tap water and sieves of 125µm and 63µm. All valves belonging to ostracods and

foraminifera were counted. Juvenile ostracods valves in the size fraction of 63–125 µm were only observed to ensure that the population is paleoenvironmental reliable (in situ assemblage). Species identification and their paleoecological interpretation were done based on Opreanu (2004), Meisch (2000) (ostracods); Kaminski et al. (2002), Briceag and Ion (2014) (foraminifers).

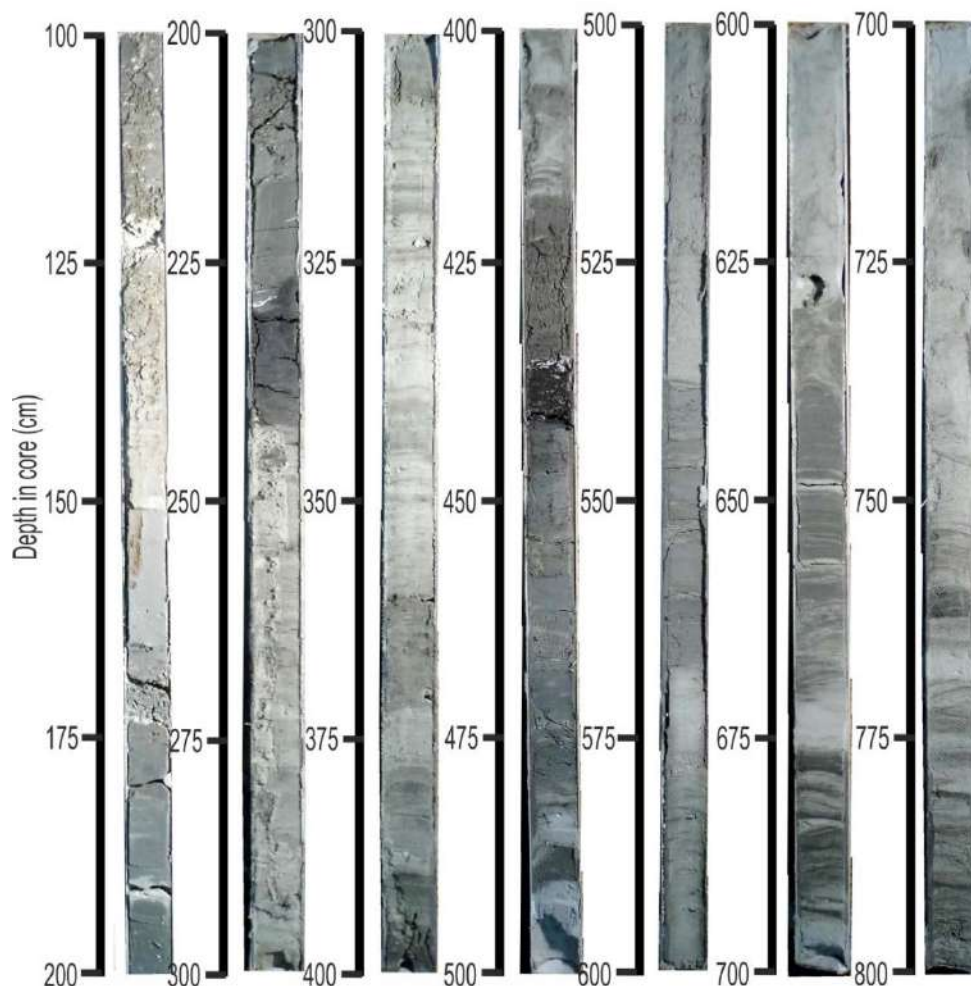


Figure 4 Photographic cores sequence shows the 1–8 meters stratigraphy

3. Results

A general stratigraphy based on visual logging, grain-size parameters (mean-size and sorting), LOI, micro-faunal and pollen analyses and magnetic susceptibility measurement is presented in Figure 3.

The core is mainly composed by poorly sorted muds and sands including vegetal remains and one thin peat layer (at 435–440 cm) which could be dated. In the first two meters distinct layers with

high concentration of bivalves and gasteropods were encountered, which after a relative absence reappear as abundant in the uppermost 40 centimetres. As a general tendency, the sand content decreases upward in the core while the clays contribution increases.

The LOI analysis is indicative of the content of organic matter (OM) and inorganic carbon (Cinorg). The OM is related to the foliar, wood fragments and

peat layers whilst Cinorg is associated with coquina, fish bones and different sediments which contain CaCO_3 . Magnetic susceptibility generally shows higher values in sandy units and smaller in silts. Regarding units, the highest average susceptibility is found in Unit 3 ($3.3\text{SI}\cdot 10^4$) and the smallest in Unit 4 ($2.3\text{SI}\cdot 10^4$), while in the first 2 units have an average of $2.6\text{SI}\cdot 10^4$.

All variations of those parameters, plus 13 samples which were analysed for micro-fauna or/and pollen content, describe the evolution model of the area divided into four stratigraphic units, in ascending order: 1 – a sandy unit with tens of layers of silts (Fig. 3, 4) which was settled before ~ 8500 years BP, 2 – homogeneous fine and very fine sand (until ~ 7900 years BP), 3 – a matrix of mostly coarse silts and very fine sands which is composed from 7 layers with different thickness from which three of them are silty (3.1, 3.3, 3.5 and 3.7) and another three are sandy (3.2; 3.4 and 3.6) (until ~ 5000 yrs BP), and unit 4 – which is composed by fine silts with a content of more than 15% clay (was present until the middle of the 20th century).

3.1 Chronology

The chronological framework is built on five ages obtained on three different type of material (Table 1). However, only four of them were used for age–depth model, because the sample from 782 cm is younger than the above sample (692cm) and con-

sidering the very small quantity of material which hampered the dating procedures we deemed it as outlier.

Regarding the age–depth model (Fig.5), this was obtained as a linear interpolation between dated levels with Clam in R-program (R $\times$ 64 3.5.1). The model shows high sedimentation rates before 5300 ± 30 BP (>1.5 mm/y), followed by a slow aggradation between 5300 ± 30 and 2491 ± 130 BP (<0.35 mm/y) with a slight increase after this date (>0.65 mm/y).

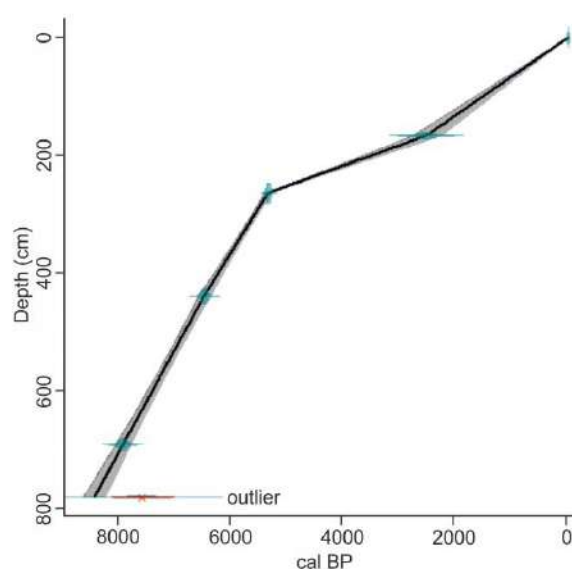


Figure 5 Age–depth model of the B1 core (obtained in R with Clam software)

Table 1 Summary of radiocarbon age results. AMS ^{14}C dating was carried out by the GADAM Centre Gliwice Absolute Dating Methods Centre (marked with asterisk) and RoAMS Laboratory of the Horia Hulubei National Institute for R&D in Physics and Nuclear Engineering. All absolute ages used in the paper were converted in calibrated years BP (with reference to 1950 AD)

Sample	Depth (m)	Material	Coordinates (lat N/long E)	Conventional ^{14}C age (years BP)	Cal. Age (BP)
Bra_62	168	Foliar	45°28'00.7"/ 28°07'25.1"	2452±37	2491±130
Bra_60*	265	Peat	45°28'00.7"/ 28°07'25.1"	4590±35	5300±30
Bra_65*	440	Foliar	45°28'00.7"/ 28°07'25.1"	5650±30	6444±52
Bra_68*	692	Wood	45°28'00.7"/ 28°07'25.1"	7060±40	7893±71
Bra_69	782	Foliar	45°28'00.7"/ 28°07'25.1"	6660±158	7552±285

3.2 Description of stratigraphic units

Unit 1 is situated between 800 to 755 centimetres. The grain-size is composed by tens of layers of silt interbedded within a poorly sorted fine sandy matrix (Fig. 4). This unit has a small content of Corg (~3%) but a high content of Cinorg (~8.4%) and a medium magnetic susceptibility.

This is the only unit with the highest weight of marine dinoflagellata species (*Alexandrium pseudogonyaulax*; *Brigantedinium simplex*; *Petapharsodinium dalei*; etc), salt water ostracodes (*Euxinocythere* sp.; *Cytheromorpha fuscata* and *Euxinocythere iopatici*) and foraminifera (*Lagena vulgaris*; *Entolinguina deplanata*) (Fig. 6). At the same time, there is a high content of degraded seeds and pollen grains besides many fragments of micro-fauna which cannot be identified. In terms of pollen, the *Pinus* and *Poaceae* species prevail against *Picea*, *Betula* and *Tilia* which are comparatively less represented (Fig. 7). Charcoal is also abundant.

Unit 2 fits 755 to 694 cm depth interval. The grain-size is composed of homogeneous fine sand, medium to poor sorted, yet better than that of Unit 1. This unit has a very small content of Corg (~1.3%) but significantly higher for Cinorg (~7.3%) whereas magnetic susceptibility maintains at similar levels as the adjacent units. The basal part of this unit is populated by a few specimens of foraminifera while the number of pollen granules increases for *Betula*, *Ulmus* and *Picea* species.

Unit 3 is the thickest unit (4.5 meters), extending from 694 to 242 cm depth. It has 7 distinct layers: four of them composed by sands and sandy silt (3.1; 3.3; 3.5; 3.7 in Fig.3) and the other three are made of silts (3.2; 3.4; 3.6). The silty layers are generally better sorted than sandy layers while mean of Corg content is almost double in the finer sub-units (4.8% vs. 2.7%, but in the sub-unit 3.5 it reaches a maximum value of 26%); mean of MS values measuring just 3.15×10^4 vs 3.7×10^4 .

Ostracodes are present only in the basal part (subunits 3.1 and 3.2) and they are mostly freshwater specimens (there are also present some brackish specimens counting for 18%); in the rest of the unit they are missing.

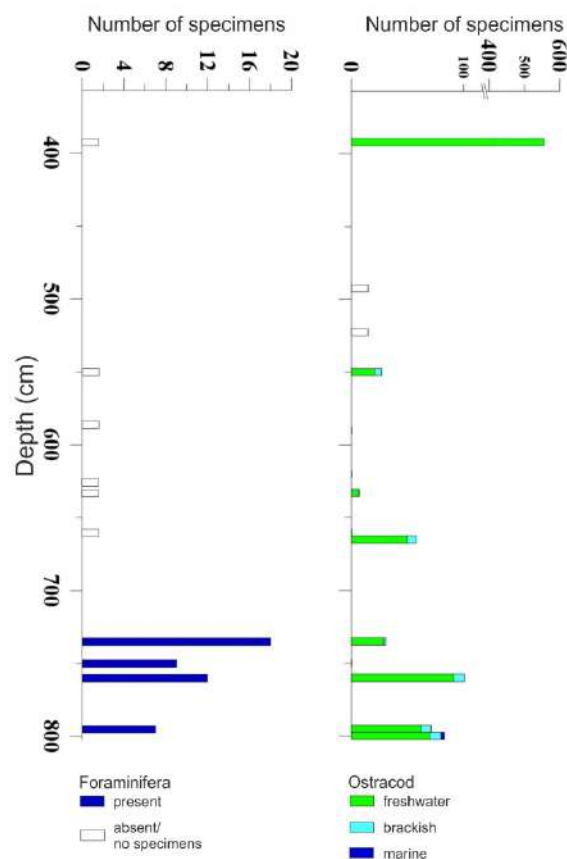
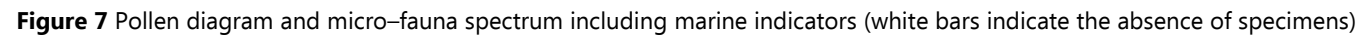


Figure 6 Ostracode and foraminifera assemblages

Unit 4 is located within the uppermost 242 centimetres, occupying the longest time period, due to a reduced sedimentation rate (Fig. 5). Grain-size is mostly fine silts and clay including few layers of mixed content (from medium sands to clay) which are very poor sorted. As an average, Unit 4 consists of more than 15% clay, 70–80% silt and approximately 5% sand, with a finer content in the middle. Corg content reaches the highest values (~ 6.9%) and Cinorg is slightly above average (7.8%). Magnetic susceptibility has the smallest mean values from the entire core with several downs below 2×10^4 , but also the main peak placed suddenly at 222 cm depth without any other transition. The degree of degradation for micro-fauna and pollen is very small compared to the basal part of the core. Grains of tree species, such as *Pinus*, *Picea*, *Betula*, *Ulmus*, have a decrease in percent, while others (*Plopus* and *Quercus*) have a slight increase or appear only here. The highest increase is among granules of the species *Polypodiaceae* (Fig. 7).



4. Discussions

After the flooding of the Black Sea by the Mediterranean waters which occurred ca. 9200 BP (Soulet et al., 2011), the present day territory of the Danube delta and lower Danube floodplain was transformed into a shallow Danube Bay of varying widths constrained by the adjacent high relief. Taking into account the previous coring (Nowacki and Wunderlich, 2012; Benecke et al., 2013) performed upstream of Bălţile Dunării (which delimits the 15–25 km wide floodplain encompassed between the Danube arms between Călăraşi and Brăila) we assume that the Danube mouth was forced upstream by sea level rise as far as Giurgiu area, whilst in front of it an open-water lakescape developed (Nowacki et al., 2018).

During 9000 – 8000 BP interval, despite of the sea level still rising at a high pace of ca. 2–4 mm/yr (Lambeck and Purcell, 2005), Danube easily advanced downstream as documented by the modern delta plain building which started to form no later than 7500 BP (Vespremeanu-Stroe et al., 2017a, b). In this context, by its intermediate position within the Lower Danube Valley Brateş Lake is a proxy for the evolution of Cotul Dunării area (the region of Danube valley turning from S–N to W–E direction) and provide the first assessment of Danube river advance towards the Black Sea within the lower Danube floodplain. This study delivers valuable information about the timing of Danube river advancement into the Black Sea.

The sedimentologic and magnetometric analyses together with different microfauna and pollen determinations performed at several key-depths led to the identification of four main units which corresponds to major phases of landscape changes which affected the regional history of the (present) Brateş Lake area during the mid- and late-Holocene.

4.1 Estuarine delta front (Unit-1): before 8000 BP

The stratigraphy of Unit 1 (Fig. 3, 4) is characterized by the alternation of fine (silts) lamina or thin layers and coarse sediments (sands) which is indicative for a delta front advance in the context of a large content of Cinorg and of a very high ratio between the Cinorg and Corg, which reach here the greatest

values for the entire core, suggesting the influence of the marine environment (Wang et al., 2011). Foraminifera were found exclusively at depths bigger than 7.3 m whilst the ostracods assemblage contains a few marine species only in the lowermost sample, at 8 m depths, but more brackish species are present in this unit at different depths with the highest density even still low (12–16%) (Fig. 6). Moreover, Unit-1 is the only one which contains different marine dinoflagellata species (Fig. 7) but also a high content of degraded seeds and pollen grains besides many fragments of micro-fauna which cannot be identified due to energetic conditions of a brackish environment. All these proxies recommend Unit-1 as being part of a shoreface unit, respectively of the subaqueous Danube delta front advancing into an estuarine-like bay.

4.2 Sandy foreshore / beach deposits (Unit-2): 8000 – 7900 BP

The presence of massive and better sorted homogeneous sands (than in Unit-1) and of some foraminifera indicate swash sediment transport. On the other hand, the relatively small thickness of this deposit (0.6 m) and the modest sorting indicate a low-wave energy environment, yet still capable of building a thin sandy beach at the medium height of a fast advancing river delta shoreline. The age of 7893 ± 71 yrs BP (Table 1) measured on the limit between the Unit-2 and Unit-3 (Fig. 2) is the minimum age of the local (Danube) mobile coastline on its way to the Black Sea.

4.3 River floodplain (Unit-3): 7900 – 5300/5000 BP

Overlain on foreshore beach sands, the Unit-3 has a 4.5 m thick stratigraphy defined by the cyclic alternation of 7 main sub-units dominated either by medium-coarse (sands) or by fine (silts) sediments noted with odd, respectively even number in Figure 3. The sandy sub-units represent the rapid aggradation phases induced by secondary channel development (for the massive sandy layers), overbank and proximal crevasse-splay deposits (for the alternating structures) which are almost sterile for microfauna.

The fine sub-units (3.2, 3.4, 3.6) are the product of the interbedding of distal parts of crevasse-splays and lacustrine sediments. They are always associated with LOI maxima (Fig. 3) and even if they are thinner than sandy sub-units, they become progressively thicker upcore (25, 30 and 60 cm) and represent the wet/submerged phases of the local floodplain when the site was covered by still shallow waters. The latest/uppermost sub-unit (3.6) is not only the largest fine sequence but it also includes the Corg main peak (at 440 cm depth) which corresponds to a peat layer covering the silty deposits which may be considered of lacustrine origin. The sandy sub-units represent the rapid aggradation phases induced by secondary channel development and overbank.

During this stage the mean sedimentation rate was of 1.6 mm/yr, respectively similar with that of the mean sea level rise (1–2 mm/yr, which slowed down over time), which reveals a complex river floodplain evolution marked by successive stages of local submergence with swamps and shallow lake inception due to water level rise and rapid sandy aggradation phases during secondary channels development and associated crevasse-splays.

4.4 Floodplain Lake (Unit-4): 5300/5000 – present

River floodplain turned into a permanent lake around 5300–5000 BP, once with the relative stabilisation of the Holocene sea level and concomitant with the formation of the first Danube delta open-coast lobes (Giosan et al., 2006; Vespremeanu-Stroe et al., 2017a). In fact, between Unit-3 and Unit-4 is a sharp contact induced by a drop in the mean grain-size and a jump of clay content and LOI, especially for Corg (Fig. 3). Moreover, the lower 10-cm layer is a dark grey layer (Fig. 4) with high organic matter content indicating an episode of a marshy wetland which subsequently transformed into a shallow lake.

The sedimentation rate computed for this unit is of ca. 0.48 mm/yr which is the minimum of the last 8000 years with the lowest rate of 0.35 mm/yr during first stage of lake development (5300–2500 BP) but significantly higher (0.65 mm/yr) since Antiquity to present in accordance with Danube sediment

load increase due to man-induced vegetation changes in the watershed (Kaplan et al., 2009).

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Recognition of erosion risk areas using Neural Network Technology: an application to the Island of Corfu

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ABSTRACT

There is a wide range of alternative approaches to study erosion processes. In this paper, we describe the construction of a model based on the interaction of Geographical Information System (GIS) and Artificial Neural Networks (ANN). The neural model uses supervised competitive learning process. The whole process begins with the digitization of collected data and the definition of the input variables, such as slope form and gradient, susceptibility to erosion and protective cover. The input variables are transformed into the erosion risk output variable using the neural model. The last stage is the development of a map of erosion risk zones. As a case study the island of Corfu (Ionian Sea, Greece) was chosen, which consists of lithologies very vulnerable to erosion and receives considerable amounts of rainfall, especially in comparison to the rest of Greece. Finally, the whole model was validated and its proper function was confirmed by field data observations.

KEYWORDS

neural networks; geomorphological process; erosion; Ionian Sea

1. Introduction

Erosion is a multistep geomorphological process. Weathering occurs before erosion. The thickness of the weathered material is interdependent of climatic variables such as temperature and precipitation, substratum rocks' susceptibility to erosion, vegetative cover and topography. The erosion process carries away the weathered material, however, there

are many factors controlling that procedure. Erosion depends on geoenvironmental factors, such as climate, rock, morphology, land uses, weathered cap, and soil management, all of which determine its intensity and magnitude (Evelpidou et al., 2018).

The identification of erosion risk zones involves a series of different stages: field work, air-photos interpretation, satellite image analysis, digitization

of geological, topographical and drainage system maps and finally the definition of input parameters. There is a vast literature concerning soil erosion or gravity movement processes with different approaches (Brundsen et al., 1975; Carrara et al., 1977; Malgot and Mahr, 1979; Ives and Messerli, 1981; Carrara, 1983; Morgan et al., 1984; Stocking et al., 1988; Carrara et al., 1991; Morgan, 1996; Marinos et al., 1997; Binagli et al., 1998; Stassopoulou et al., 1998; Gournelos et al., 2004).

Artificial Neural Networks (ANN) are characterized by a number of connected processing units, which have the ability to adapt their structure to multiple input data. The capabilities of learning and generalizing are an important asset for classification purposes. There are diverse architectures of ANN and different learning algorithms (Haykin, 1994). Artificial neural systems have been applied to a series of problems related to geosciences have shown a great potential. Some representative examples are those from multisource and hyperspectral data classification (Benediktsson et al., 1990), modelling of radioactive contaminated territories (Wu and Zhou, 1993), spatial simulation (Dowd, 1994), remote sensing imagery classification (Paola and Schowengerdt, 1995) and other similar applications (Kanevski et al., 1996; Wong et al., 2002; Licznar and Nearing, 2003). Recently similar researches have been applied for geotechnical and geology engineering (Ren et al., 2018; Arabi Aliabad et al., 2018), developing models for air pollution mining management (Asif et al., 2018), identifying rock strength correlations (Tariq et al., 2017). ANNs have also been employed for landslides susceptibility mapping (e.g. Yesilnacar and Topal, 2005; Melchiorre et al., 2008) and soil erosion studies (e.g. De la Rosa et al., 1999; Behrens et al., 2005).

In this context, the aim of this paper is to determine erosion risk areas on the island of Corfu, using Neural Network Technology. A simple competitive network, with four input and four competitive neurons, was constructed. The output of the system is a function of input vector, the weight matrix and the bias vector. Competitive learning processes are the outgrowth of the learning subspace method (LSM) (Kohonen, 1991), where the winner

subspace is the closest of a given subspace. In its simplest form, a competitive network is made of a number of neurons, which receive the same vectorial data and by mutual interaction, one of the neurons wins (Kohonen, 1982; Rumelhart and Zipser, 1985). The election of the winner neuron is achieved by comparing the dot products of input vectors and the connections. The neuron with the highest value of dot product is the winner. When the inputs are normalised, they may be represented in the unit hypersphere. The learning procedure consists of progressive rotation of the weight vector to the closest cluster of input data (Gurney, 1997).

ANNs are computational structures, which can solve complex problems and currently are among the most successful algorithms for detection and classification of stream data (Ghorbani et al., 2018). In the present work, the definition of recognition zones was not very straightforward, due to the difficulties involved in defining precise values for the input variables.

2. Study area

Corfu is the second largest Ionian Island (Figs. 1, 2), covering an area of 583.4 km², with a length of 61.65 km and a width of 26.58 km (maximum width). Geologically the island belongs to the external Ionian isopic zone as defined by Aubouin and Der-court (1962), Caputo (1984), Underhill (1989) and Doutsos and Frydas (1994). The alpine structure mainly includes a Mesozoic carbonate platform, the progressive fracturing and deepening of which, is reflected on the pelagic sedimentation, followed by molassic sediments and Pliocene and Quaternary deposits (Fig. 3).

The climate of the island is Mediterranean, with cool summers and mild winters. However, its main characteristic is the high amount of rainfalls (mean annual precipitation more than 1.180 mm (Kandilis, 1991) (Fig. 4, Table 1). The climatic conditions in combination to the lithology of the island have contributed to the development, at some areas, of fertile soil layers with a rich vegetative cover, such as olives, vines, oranges, lemons, maize, cypress, pines and figs.



Figure 1 Location map of study area

Three geomorphological units may be distinguished on the island of Corfu:

1. The northern part, which includes the eastern part of Pantokrator limestone mountain (906 m) and the western part, which includes a sedimentary basin with Neogene sediments,
2. The central part, with smooth hilly landforms composed of Neogene and gypsum formations,
3. The southern part, with a low lying topography, and dominating lithologies of Neogene and Quaternary formations (Fig. 3).

The drainage system of Corfu is mainly characterised by a number of small streams, draining the Neogene formations, in the northern part, and the gypsum and Neogene sediments, in the central and southern part (Fig. 3).

On Corfu, apart from the marine and fluvial processes, the aeolian processes are also influential; this is indicated by the development of small dunes, especially along the banks of the Korisia Lake (Leontaris, 1977).

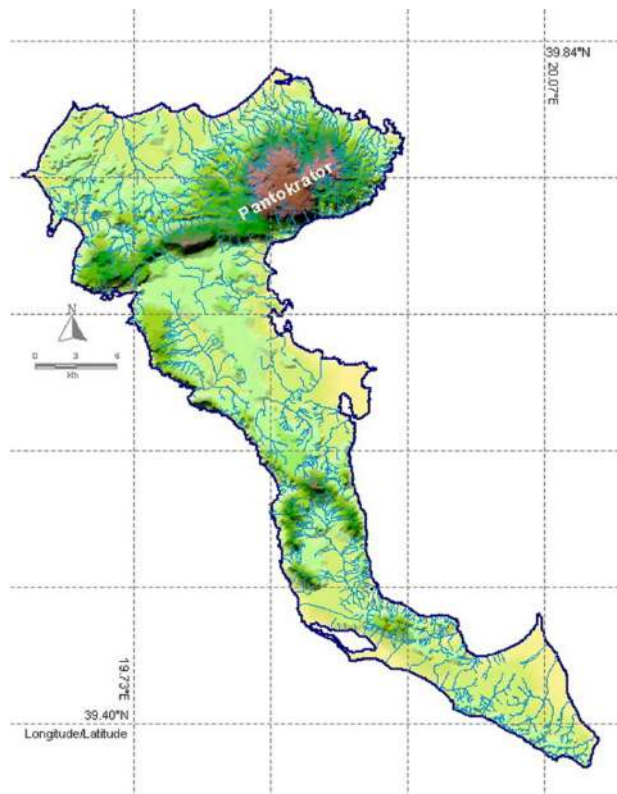


Figure 2 Topography and the drainage system of Corfu Island

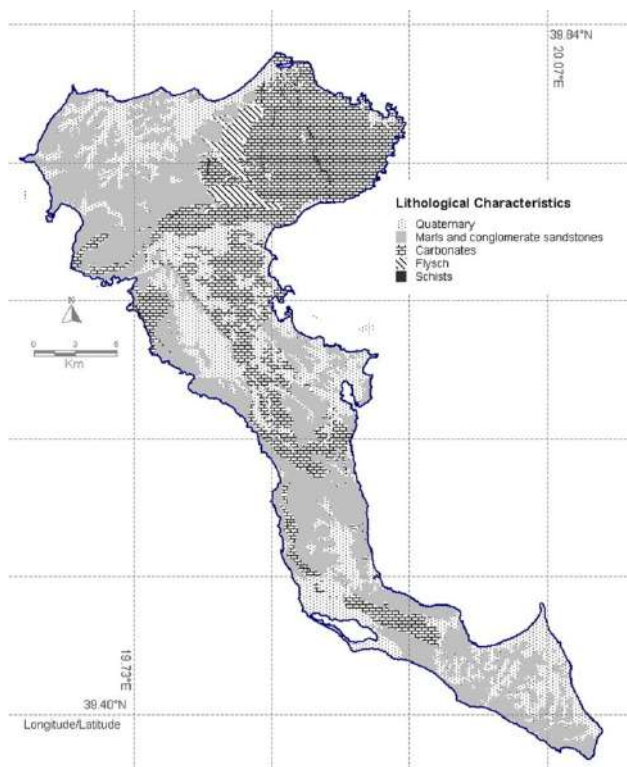


Figure 3 Lithological map of Corfu Island

3. Methodology

The methodology of this study was divided in several steps (Fig. 5). The process of developing a map of risk zones has been approached by using a loose coupling of GIS and Matlab softwares. The first step included the collection of data, which derived from existing geological maps (Institute of Geology and Mineral Exploration, 1970), topographic maps in scale 1:50 000 (Hellenic Army Geographical Service 1969, 1980), aerial photos in scale 1:33 000 and satellite images (Landsat 1980). Field work was also accomplished, in order to collect additional data, understand the processes taking place and select the appropriate input variables for the study area (geomorphology, slope gradient, vulnerability of erosion, protective cover).

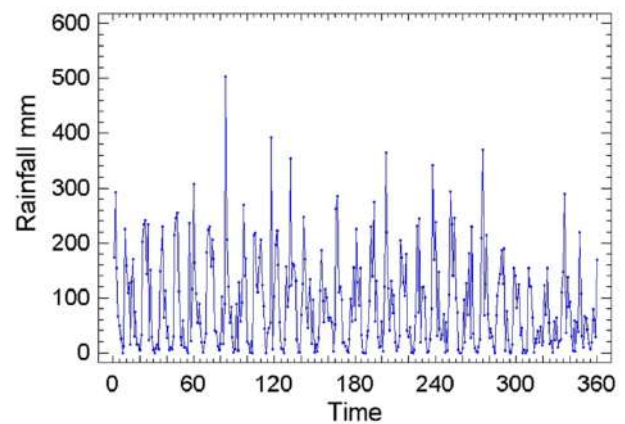


Figure 4 Time series plot for rainfall monthly data of Corfu

Table 1 Summary statistics of 30 years (1963–1992) rainfall in mm in Corfu island

Count	360
Average	87.3167
Variance	7377.94
Standard deviation	85.8949
Minimum	0.0
Range	503.7
Standard skewness	9.86789
Standard kurtosis	6.82953

The next step included the development of a GIS database, through the digitization of existing data. The input variables were processed by using a

neural network system for studying the erosional – depositional processes. Initially, the spatial database of all input variables has been developed in a GIS platform (MapInfo) and subsequently it was analysed, which led to the production of thematic maps. Values of the input variables were estimated for each elementary cell (a grid based database). The output variable at this final stage was erosion risk. In order to combine data layers, all the original data have been normalised by dividing them by their maximum value.

The input variables were applied to Matlab, after their normalisation, where the architecture of the ANN was developed. In order for an ANN to classify input vectors it must first be initialised, and then trained. The initialisation processes attribute values to the weight matrix. The training phase aims to the

progressive minimisation of the distance between the winning neuron and the input vector.

Afterwards, a sensitivity analysis was carried out in order to test the proposed model to small perturbation of the input values. The analysis confirmed that minimal changes had no significant effect on the output variable. Following, the output classes of the ANN were imported in the GIS in order to produce the final output map. In this manner, the spatial capabilities of GIS and the classification power of the ANN system were fully exploited. It should be noted that, although the ANNs are considered to be black – box models, in this research the transparency of all the procedures and the continuous control of the results through field observations minimised this weakness.

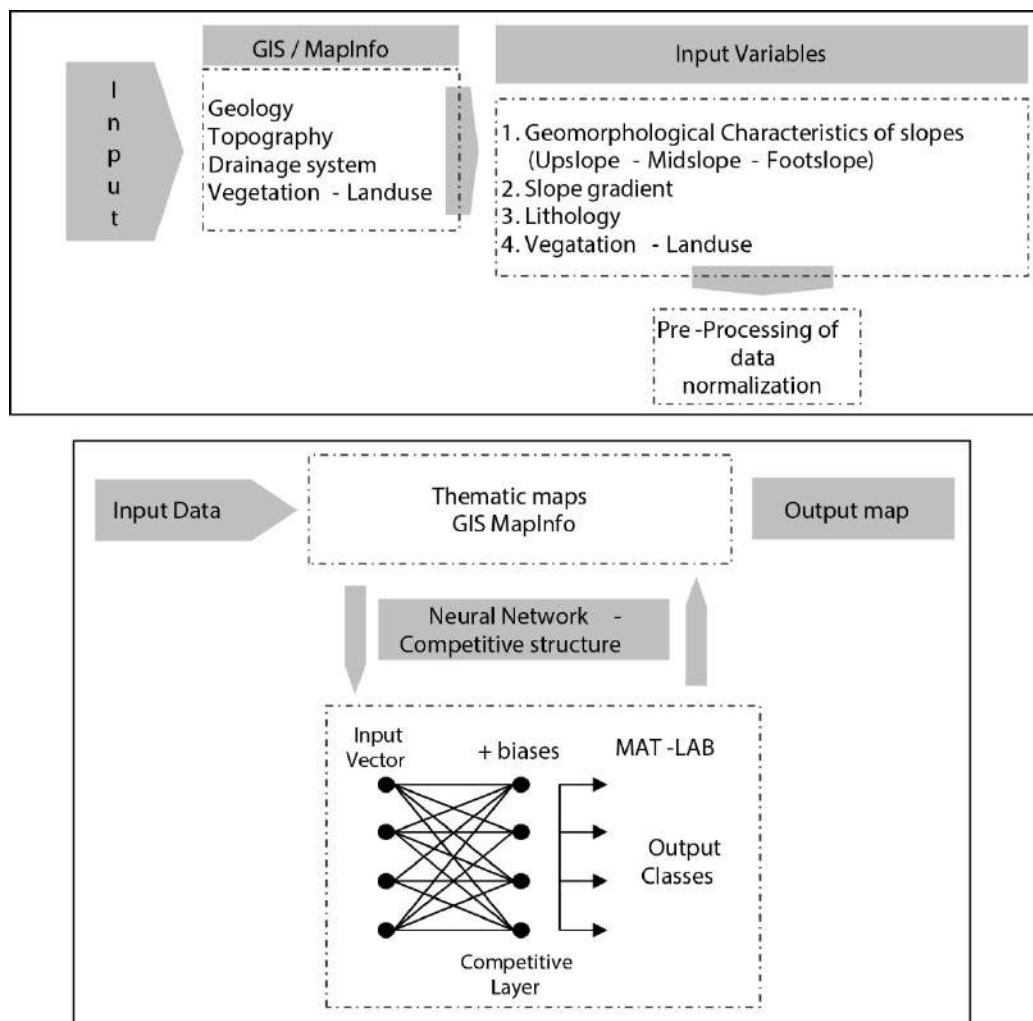


Figure 5 The methodological steps to produce the final output map: the erosion risk map

4. Results and Discussion

The slope characteristics are influential. The geomorphology of slopes depends on the slope form, whose elements reflect the exogenous processes involved, such as weathering and erosion. A typical three division of a slope's elements has been adopted: upslope, mid-slope and footslope by

scaling their property of favouring erosion processes. Upslope is confirmed as an area of relatively low erosion (0.2–0.3), mid-slope is dominated by high sediment transport (0.7–1) and footslope is mainly an area of deposition and therefore a very low degree of erosion is taking place (Table 2).

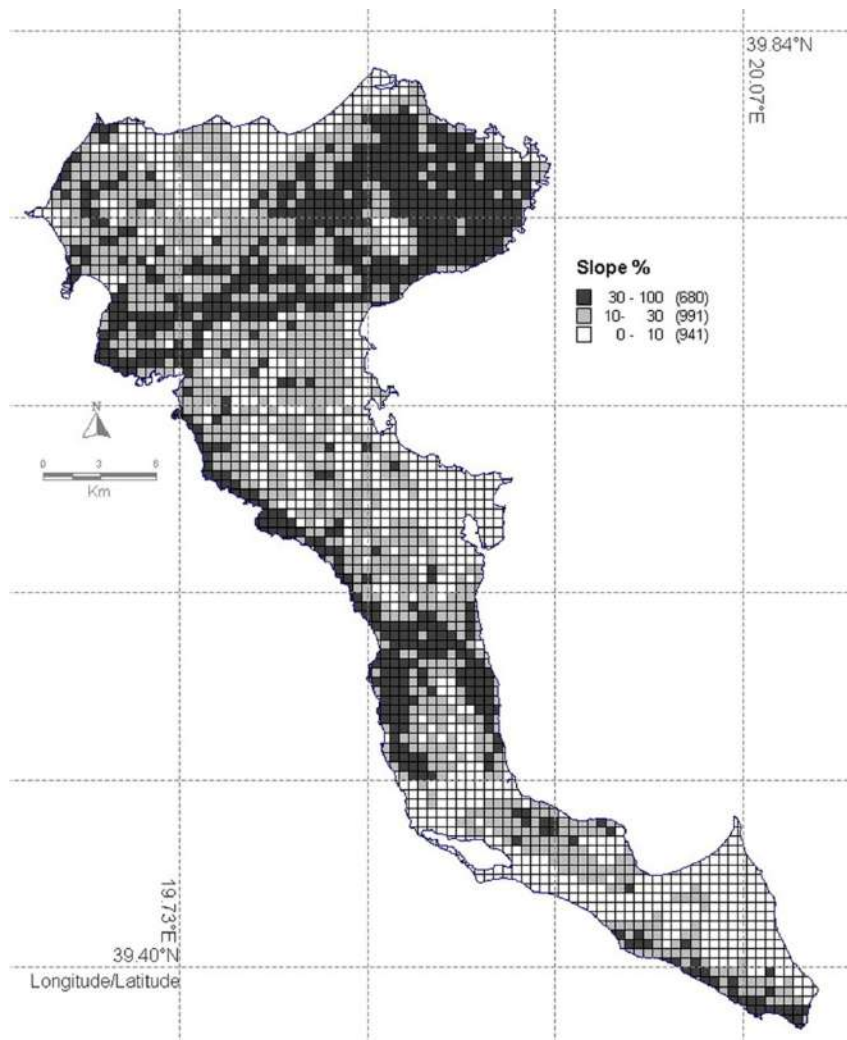


Figure 6 Slope gradient map of the Island of Corfu

The second processed variable was the morphological slope gradient. It is obvious that the slope steepness is critical for the intensity of erosion. Sediment removal is estimated as a function of divide distance multiplied by a power of slope gradient (Kirkby, 1978). Where slopes are steeper, gravitational movement and downslope sediment delivery are higher. In general, a linear or power relation between slope gradient and erosional

products has been noted (Kirkby, 1969; Schumm, 1977). Slope steepness is critical for the intensity of erosion. Therefore, slope gradient was used as the second variable, and was calculated through the digital elevation model derived from GIS analysis. Three slope gradient classes were distinguished based on their steepness: high (0.7–1), medium (0.4–0.7) and low (0.1–0.4) (Fig. 6), (Table 2).

Susceptibility to erosion is an important factor controlling the hardness of rocks, while the combination of rock's composition and texture, expressed by physical resistance, also affects erosion rates (Meybeck, 1987). The third variable, rock's susceptibility to erosion, was distinguished in three classes based on field observations and literature (Kuenen, 1956; Jensen and Painter, 1974; Selby, 1987). This variable is difficult to measure directly. Observations on the rocks' resistance to abrasion have resulted to a list of rocks sorted by descending order according to their resistance to erosion (Kuenen, 1956). On the

other hand, Selby (1987) has proposed a rock mass strength classification and rating, to express the resistance to erosion. In the aforementioned classification limestones are more resistant to erosion than schists. The first attempt to assign susceptibility to erosion values on different rock types was made by Jensen and Painter (1974). For Corfu Island, low values for limestones (0.1–0.3), medium values for the schists (0.3–0.6) and high values for the Neogene and Quaternary sediments (0.6–1), were adopted (Table 2).

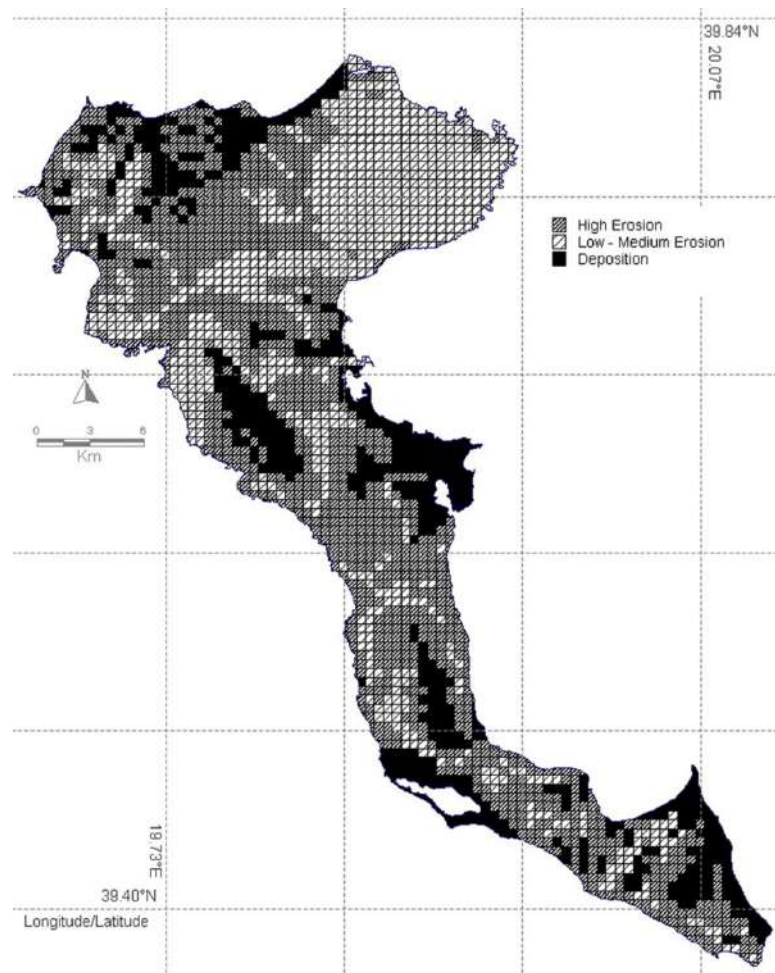


Figure 7 Erosion risk map of the Island of Corfu

The protective cover is closely linked to the soil characteristics and consequently to the ability of water for infiltration and run off. The protective cover variable was estimated mainly through data from vegetative cover, which protects the weathering material, by reducing the energy of the rain,

controlling the soil thickness and the weathering mantle, increasing the infiltration capacity and limiting runoff. The vegetative cover of Corfu Island was divided in three classes: trees and bushes, agricultural lands and barren rocks.

Table 2 Distribution values of the input and the output data sets

Input Variables				Output Variables		
Type	Range	Count	Ratio %	Erosion Risk	Count	Ratio %
Slope Form	Upslope 0.2–0.3	481	18.417	High	910	34.839
	Midslope 0.7–1	1515	58	Medium – Low	1059	40.544
	Footslope 0–0.1	616	23.583	Deposition	643	24.617
Slope Gradient	High 30–100%	5	0.19			
	0.7–1					
	Medium 10–30%	261	9.994			
	0.4–0.7					
Lithology	Low 0–10%	2346	89.816			
	0.1–0.4					
	High to erosion 0.6–1	1696	64.932			
Vegetation Landuse	Medium to erosion 0.3–0.6	385	14.739			
	Low to erosion 0.1–0.3	531	20.329			
	Trees and Bushes 0.1–0.2	868	33.232			
Vegetation Landuse	Agricultural land 0.2–0.7	1645	62.978			
	Barren rocks 0.7–1	99	3.79			

These classes have been empirically valued according to their contribution to the erosional processes, as follows: 0.1–0.2 for trees and bushes, 0.2–0.7 for agricultural lands and 0.7–1 for barren rocks (Table 2).

The four input variables using the ANN model were transformed into the output variable, which was the erosion risk (Figs. 5, 7). In the map of Figure 7 three classes of geomorphological processes may be distinguished along with three zones of high erosion, medium and low erosion and deposition.

It is obvious that high erosion zones are located where very susceptible to erosion rocks are present (Neogene or Quaternary deposits), slope gradients are relatively high and mid-slopes exist. Low to medium erosion zones are found in areas of schists or

limestones, with medium slope gradients and mainly on upslopes or in mid-slopes.

Finally, deposition zones are restricted to foot-slope areas. The aforementioned results have been further confirmed through field observations on Corfu Island.

5. Conclusions

Artificial Neural Networks and GIS may be combined to produce erosion risk maps. The implementation of these procedures has been achieved by loose coupling of the GIS platform (MapInfo) and a constructed ANN (Mat-Lab). This application was used to map the erosion risk zones of Corfu Island. Our results show that high erosion risk areas are located on the northwestern and central part of the

island, while low to medium risk areas are mainly located in the northeastern part. Depositional zones are restricted mainly in coastal areas. The analysis of the zones of high erosion risk may result to the recognition of areas where land degradation processes are very active and highlight areas, where measures are needed against further soil loss and superficial deposits. This type of model is relatively easy to develop, is rapidly adapted to input data and is very easy to generalize.

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Simplifying a deltaic labyrinth: anthropogenic imprint on river deltas

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ABSTRACT

The information contained by historical maps provides a good source of understanding the complex transformation of a deltaic environment by human activity. Using the Danube delta as an example, here we show that a cartographic diagnosis for river deltas is based on four main steps that outline the learning stages for every similar area: 1) exploring coasts (for the early stages of the portolan and Ptolemaic maps); 2) exploring depths (for the succeeding imperial and military maps which focused on the access along the deltaic distributaries); 3) exploring deltaic networks (when economic and ecological reasons led to detailed topographic maps based on field measurements and aerial photos); 4) ecological protectionism (when ecological reasons dictate land use patterns and determine land use change). This synopsis is applicable to other river deltas with some adaptations imposed by the local context. We interpret the four stages in the description of the Delta as resting on and further reinforcing the power of the centre to dictate the uses of the periphery. We further argue that the way the territory is lived by local inhabitants is continuously marginalized and effaced. This stands in the way of future adaptive strategies.

KEYWORDS

Danube delta; anthropogenic imprint; Black Sea; river deltas; human ecology; Anthropocene

1. Introduction

Sedimentary, paleoenvironmental and paleogenetic evidence shows us that land use occurred in both the Danube Delta and the Black Sea basin well before the Industrial Revolution started (Giosan et al., 2012). The information about the anthropogenic

footprint contained by maps can play a considerable role in understanding the alteration of the environment, highlighting the changes produced in the last centuries. The human impact on the environment is also reflected by the relationship between mapping and governance which is both pervasive and power-

ful (Akerman, 2009). Maps always served as a primary technology of administration that governments used to extend their authority and control new territories or grab land (Edney, 2009). They can also be interpreted as fundamental justifications for later activities in the charted territories, means by which power extends outwards from the centre to the periphery. As we will see, mapping the Delta has gone hand in hand with transforming it to better fit a flat and rational approach to the territory.

The map in this view is not a benign representation of the territory, but rather one crucial aspect of the wielding of power (Elden, 2013; Steinberg, 2009). Therefore, the changes we trace in cartographical production reflect both changes to the territory and to the wielding of power in that territory. Studying the progressive changes in maps together with the progressive transformation of the Delta and its uses shows different stages of knowledge corresponding to different stages of use and control (see Figure 2). This is so in two senses: what the maps do not show becomes part of the justification for later activities in the territory, and changes to the territory are mostly achieved through the dominance of a cartographical mentality. In other words, portraying the Delta as empty (so not showing settlements) allows greater access to it, and remaking the physical Delta into a flatter space starts with the engineering exercise of plotting what one wishes were the case. These two aspects of cartographical knowledge are here shown inseparable.

Population and economic growth had a profound impact on deltas (Vörösmarty et al., 2009; Syvitski et al., 2009). Similarly, human engineering has had a considerable influence on the evolution of many wetlands (Syvitski and Saito, 2007; Giosan et al., 2013). However, wetlands have not traditionally been easy to chart, accessibility having been the main problem for a cartographic approach (the shallow water and marshes act as natural barriers). Furthermore, cartographical approaches have not been particularly well equipped to deal with the dynamism of a deltaic environment: neither fully water nor solid land, a delta is an environment in-between that constantly changes and frustrates attempts to define and draw it narrowly. It is there-

fore best understood as a dynamic volume (also see Steinberg and Peters, 2015).

Consequently, a deltaic space has always been a refuge for political castaways and a protective barrier between empires. The Danube Delta and its floodplain are the best European example of a buffer zone between the Ottoman, Russian or Austro-Hungarian Empires (Fig. 1). This was a space with ethnic and religious diversity, with populations of Romanians, Russians, Turks, Germans, and Bulgarians (Enăchescu, 2013). The frequent changes of borders, following the frequent conflicts between Christendom and the Islamic world, involved an increasing cartographical effort for collecting data. From the 18th and 19th centuries, Austrian or Russian officers undertook new topographical surveys with a stress on accurate mathematical precision. The changes of borders can themselves be interpreted as occasioned by the Delta's fundamental liquidity – a space of flows and volumes that has always challenged the cartographical attempt of laying out a flat territory. The new cartographical results became highly important in time of peace, when the economic powers exercised their supremacy and used the Danube River as a principal route for goods transport (Constantinescu et al., 2010).

1.1 Aims and Scope of the article

Using a large cartographical database, complemented with interdisciplinary methods (see below), we develop a four-stage evolution in the knowledge of the Danube delta, applicable to other deltas as well. Through the four stages, we show the anthropogenic imprint on river deltas advancing together with changing uses of the territory and changing conceptualizations of it from the point of view of the centre of power. In other words, we will show that the Danube Delta has gone through four different phases of cartographic knowledge, each interpreted as both occasioning and being occasioned by the changing uses of the territory. The transportation of goods was aided by the charting of the main branches, but also led to their eventual shortening and the maintenance of artificial depths. This increasing access to the interior of the Delta eventually led to the cartographic exploration of the network of channels and lakes at the heart of the

territory, which allowed greater exploitation of fish and reed resources. In turn, these activities fed a transformation of the deltaic labyrinth into a navigable, accessible, eminently mappable space. The stages we propose arrive to today's ecological protectionism, which we will show to be inscribed in the same logic of cartographical knowledge understood as wielding power and 'taming' previously unknown lands.



Figure 1 Danube delta

2. Study area

Crossing ten countries and collecting its water from 19 others, the Danube is the most international river on Earth, flowing over 2,870 km from the Black Forest Mountains (Germany) to the Black Sea (Romania). The average water discharge is 6,470 m³/s and the sediment discharge is 1,555 kg/s with a catchment area covering 817,000 km² (McCarney–Castle et al., 2012).

The extent of anthropic impact in the Danube Delta was largely decided at the beginning of the 20th century, when two different models were envisioned for the Danube floodplain: an engineering model proposed to drain the floodplain for agriculture (Ionescu–Sisești, 1933), and a natural solution suggested the development of fish farms and ani-

mal husbandry while preserving the system of lakes and marshes (Antipa, 1914). Though Antipa's model managed to initially stave off massive engineering interventions in the Delta, the first model was implemented on a large scale after 1950, including now 56 embanked enclosures (Constantinescu et al., 2015). The effect on the water regime was significant: the artificial levees produced a narrowing of the stream bed with an increase in flood events, lateral or incision erosion.

The anthropogenic intervention in the Danube basin is significant, with 56 reservoirs along the river and more than 300 in the catchment area after 1945, of which 50% exist in Romania (Stanciu et al., 2008). Although the energy production is considered significant, the overlooked effect of these dams is the 60% decrease of the sediment supply transported by the river (McCarney–Castle et al., 2012). In Romania, the length of the Danube course is 993.5 km and the floodplain covers 573,000 ha, of which 75% is embanked (Constantinescu et al., 2015). The river also provides over 60% of the entire runoff reaching the Black Sea generating a strong coupling between fluvial fluxes and the marine biogeochemistry and ecology (Kideys, 2002; Syvitski, 2005; Giosan et al., 2012).

3. Data and Methods

Although a large number of materials were produced before the 19th century, they were made without a cartographic projection and coordinates. Those maps reflect just a general aspect of the Delta, focusing only on the main navigable arms. Accurate documents were published from the 19th century onward, with precise measurements, after the Crimean War: Russian maps (1830, 1854, 1856), Manganari (1841), Spratt (1856), E.C.D. (1867, 1874, 1887, 1902), the General Staff of the Army (1884), Vidrașcu (1911). After the Second World War, during the Communist regime, major topographical surveys were conducted (1960, 1980) using a particular projection used in the Soviet states, Gauss–Kruger.

In our previous papers (Constantinescu et al., 2010; Constantinescu, 2014) we investigated cartographical products regarding the Danube Delta, indexing maps and analysing the topographical techniques involved. For the present study, we

chose only maps that constituted original contributions, created from topographical field surveys, in order to be integrated into a GIS database:

- Maps belonging to the European Commission of the Danube (1856, 1886, 1926);
- General Staff of the Army (Harta Marelui Stat Major: 1880–1884) and derived maps like Planurile Directoare de Tragere (1920s–1930s);
- Vidrașcu map from 1909–1911;
- Danube navigational charts (1934–1935);
- Topographic maps of Danube Delta (1960, 1980);
- Soviet topographic maps (1980).

Georeferencing was made in the native cartographic projection (Mercator, Lambert–Cholesky, Gauss–Kruger, and UTM) with the transformation of the national grid, currently called Stereo–70. All the maps were digitised in QGIS 2.18 extracting arms and lakes, natural and artificial channels, levees, permanent and temporary settlements. Starting with Landsat missions, the maps' production took a different approach, with two images per month showing the Delta at an unprecedented temporal resolution. Later the Romanian authorities (ANCP) created a national aerial mosaic at a spatial resolution of 0.5 m (2005–2012). The last product was used to extract the same layers, to create a present-day image of the entire delta, a century after the Vidrașcu map. Finally, the results were integrated into a PostGIS database and compared in order to understand the anthropogenic imprint on this territory. Derived products which we called mirror maps were created later, like those in Figure 4.

This map archive was complemented by interviews with locals of Sfântu Gheorghe village and participant observation spanning three years, from 2014 until 2017. Historical sources detailing the uses of the deltaic space in different time periods were also consulted (Dorondel, 2005; Van Assche et al., 2009; Van Assche et al., 2011a, 2011b; Van Assche et al., 2012; Iordachi and Van Assche, 2015; von Hardenberg, 2017; Gătejel 2016, 2017; Tănăsescu, 2017). Lastly, the current legal framework applicable to the Danube Delta was studied in order to accurately define the current period of ecological protectionism (also see Dorondel and Mitroi, 2017; Prelz Oltramonti and Tănăsescu, 2018). This legal

study was complemented by the qualitative data mentioned above.

Starting from the information contained in old maps (former lakes or natural channels) a new topographical map of the delta was created (see supplementary material). This new cartographical product can offer an image of the deltaic space that includes human-induced environments but does not exclude the original natural landscape. Some toponyms have survived in local communities reminding us that a former lake, forest or creek existed in that space and our map brings them back into the present. Another electronic version was created and published using Mapbox and TileMill, available at <http://tinyurl.com/zrkhcno>. This map visualizes the cumulative dynamism of the Delta, showing both historical baselines and current use. The juxtaposition of natural state and anthropic impact can start giving an idea of the extent of change that we conceptualize and explain throughout this paper.

4. Results and discussion

Following the anthropogenic imprint in this inaccessible territory, different stages of knowledge result each corresponding to different modes of wielding power in the mapped space. Based on the cartographical evidence, there are four main phases that define the learning stages which can describe every deltaic area (see Figure 2):

- 1) Exploring coasts: for the early stages of the portolan charts and Ptolemaic maps.
- 2) Exploring depths: for the succeeding imperial and military maps which focused on the access along the deltaic distributaries.
- 3) Exploring deltaic networks: when economic and ecological reasons led to detailed topographic maps based on field measurements and aerial photos.
- 4) Ecological Protectionism: after the Danube Delta received a Biosphere Reserve status (1991).

From Antiquity to the 18th century, a descriptive and pictorial style persisted in the case of portolan charts. The technological capacity of using the deltaic territory at the time being very limited, it conditioned the ways in which the Delta could be known. It was therefore more inscribed within a symbolic order that nonetheless served the purpose of integrating the relatively unknown territory into the

known world order. The development of mathematical cartography allowed for the production of new topographic surveys based on military needs. The technological advancements of mathematical cartography worked together with the expansion of naval power to inscribe the territory of the Delta within first a military, and then an economic phase. The most important data in this period is bathymetry, precisely because the exploration of the territory and its use go hand in hand. Finally, the networks in the interior of the Delta are mapped and these serve a deepening of economic exploitation before a protectionist approach settles in. These phases of knowledge, reflected and reflective of uses of the territory and allied technologies, seem very different

from each other. However, besides being connected by the enabling role of the map in the exercise of power, they are also connected by the assumption of the deltaic space as fundamentally empty of an important local perspective. The stages of cartographical exploration reproduce a familiar dynamic of a central power exploring peripheral spaces, a process which consolidates the grip of the centre over the marginal (but highly important) land. Importantly, in describing the land as fundamentally empty the terrain is prepared for later interventions aimed at flattening the complex deltaic labyrinth and making its resources more available. In this process the map is the supreme means.

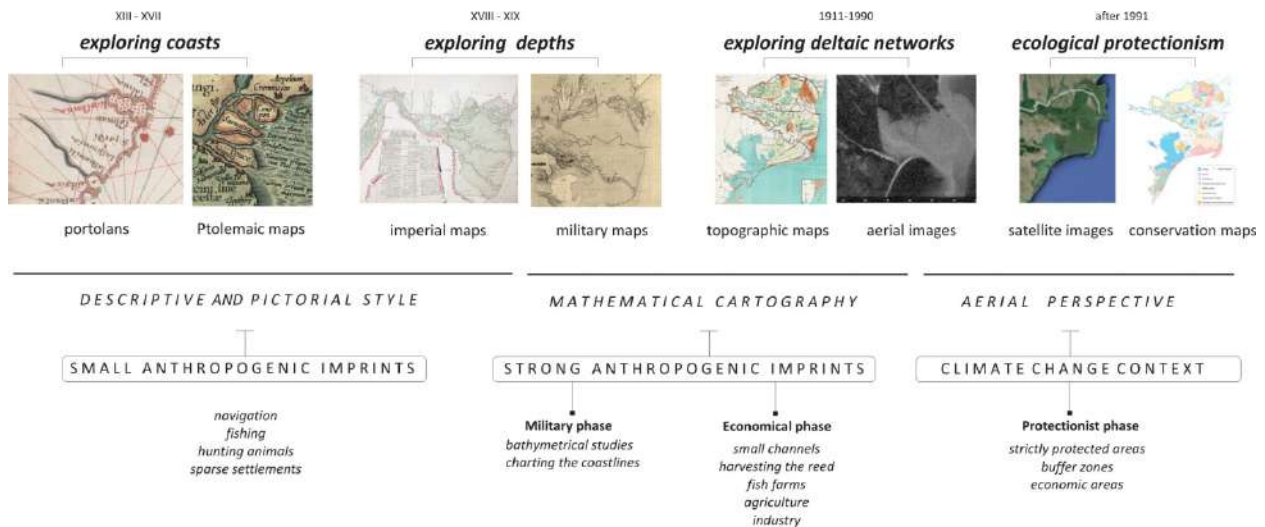


Figure 2 The scenario of cartographical knowledge regarding the Danube Delta

4.1 The Descriptive and Pictorial Style

Historical texts can yield enough information about coastline configuration, distances between river mouths or the time required for traveling between two cities. Before the appearance of the portolan charts a symbolic image of the world persisted, characterized by a profound Christian influence. The most prominent were the examples of the *mappa mundi* or *imago mundi*. From the 13th century, the new cartographic products (i.e. portolans) created an accurate shorelines' representation across the Mediterranean and the Black Sea basins. The deltaic context is usually reduced to the river mouths and coastline, with a space between the arms. In the

16th–17th centuries, there was a combination of old information (Greek or Roman sources based on a Ptolemaic representation) with accurate descriptions of travellers or military officers, yet insufficient to create reliable maps. In other words, the space was not rendered economically accessible by this stage of cartographic knowledge. Rather, it was inscribed in an established symbolic order, therefore aiding the continuation of this order. The decorative purpose in this stage is clear, and this resulted in the confusion which persisted in either the positioning of the settlements or the river mouths or islands' names, especially in an exotic part of *Mare Maggiore*. The presence of the Venetian or the Genovese ships is illustrated on maps with the flags of their cities. Since the most frequent activity was trading

with cabotage navigation, these vessels primarily undertook the coasts' exploration. However, there is a persistent lack of information about the territory between the Delta's main branches and this makes the landscape mysterious and unfit for detailed exploration.

Nonetheless, several lakes or creeks are specified, meaning that the space between arms was unreachable or did not present military interest. The navigation from the Black Sea to the upstream commercial ports on the Danube was achieved along the main arms, Sulina being the most used, which is why bathymetry was the most sought-after information that a map could provide at this time.

Descriptive boxes complete the graphical support with a focus on navigational issues (depth, dangerous points, and mouth-bar conditions). Huge empty spaces, between the main arms, suggest a no man's land sort of territory. Moreover, the anthropogenic footprints can be distinguished along the river, where the sparse settlements were offering a vague insight into an unknown land, dominated by wilderness. Fishing, reed harvesting and hunting were the only forms of anthropogenic pressure, focused particularly on the main arms, where there were just a few isolated villages and a single town, Sulina. That period could be regarded as one characterized by a type of human activity that did not fundamentally interfere with natural processes. As we know from other areas that were primarily thought of as wilderness (see Ter Steege et al., 2013, Levis et al., 2017), such as the Amazon basin, the habitation of biodiverse and rich ecological spaces likely results in their transformation. The question is whether this transformation was overall benign for ecological health or not. However, in this early period of cartographical exploration human habitation was supposed insignificant, if supposed at all.

4.2 Mathematical Cartography

At the crossroad of three empires (the Ottoman, the Russian and the Austro-Hungarian one), the Danube Delta was a marginal territory with frequent changes of borders. However, this political instability presented some cartographical advantages: Austrian and Russian officers created detailed sur-

veys, and focused their interest on the Danube's arms, where they took water depth measurements. The most outstanding product is a Russian map from 1771 with a German version (1778) and a French edition at the end of the 18th century. Depth measurements were a crucial precursor to a new kind of exploration of the Delta's interior. The map of the Delta is slowly beginning to aid physical access to the heart of the territory.

The political consequences after the Crimean War led to a new European organisation, responsible for maintaining river navigation: the European Commission of the Danube (ECD). It was a flourishing period of cartography, with new products published in several atlases and many studies, which were later kept in the archives (Constantinescu et al., 2010; Constantinescu, 2014; Budileanu, 2013). A first product, developed by Captain Thomas Abel Spratt in 1856 was based on accurate topographical surveys, without going inland into the marshes, showing the familiar image dominated by three arms and a few secondary courses.

The main anthropogenic pressure in this period was exercised along Sulina channel where a jetty structure was built, first at the river mouth, then followed by the shortening and the dredging of the river upstream (Fig. 3A). This was followed by the dense bathymetrical points measurements a few miles offshore, coastal morphometry having the same importance as the fluvial one. Between 1880 and 1902, the total length of the sectors dredged on Sulina channel was 31.9 km, accumulating a volume of 24 Mm³. The entire maritime channel must have a minimum depth of 24 feet (7.32 m). This first engineering intervention in the Delta marks a change from cartographic knowledge of a fundamentally unknown territory to the active creation of territories that would neatly fit a navigational map. The Sulina arm is shortened and its depth kept a predictable constant, and this is a radical new step in literally making a territory that renders itself accessible to the activities that the powers of the time saw fit (also see Cioc, 2002; Blackbourn, 2011; Pritchard, 2011), in this case mostly trade and navigation.

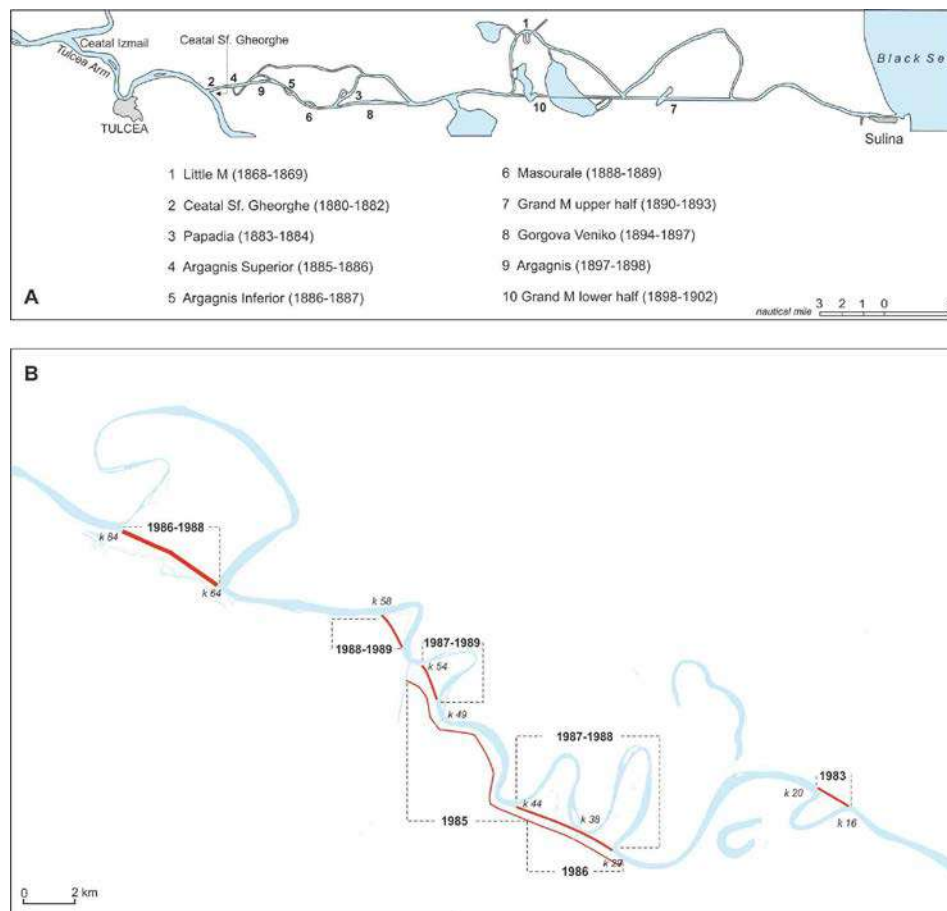


Figure 3 A. Shortening Sulina arm and transforming into a maritime navigational channel (after Rosetti and Rey, 1931);
B. New channels cut along Sf. Gheorghe arm based on Landsat images

From the beginning of the 20th century, the focus shifted to fishing activities, as the Romanian authorities sought economic benefits and territorial consolidation. For the Romanian state the new, unknown, territory represented a great possibility for natural resource extraction. Grigore Antipa was tasked in 1893 with devising fisheries in the Danube Delta in order to exploit existing stocks. By 1895, he drafted a study of Romanian fisheries (Antipa, 1895) that set the basis for the 1896 fishing law that institutionalizes Antipa's recommendations for the rational exploitation of fish stocks (Dorondel and Mitroi, 2017). He sought to increase fish production through paying close attention to local hydrological conditions and modifying the deltaic space to increase beneficial conditions and mitigate problematic ones. Under his leadership, the state sought to reverse the silting of lakes, a natural process that leads to the deterioration of fish stocks, by dredging channels and cutting new ones.

New channels were cut along the natural river course, shortening the distances and increasing the water depth for bigger boats: Dunavăț (1907), Enisala (1913), Dranov (1914), Litcov (1928–1932), Crasnicol (1930–1934). The purpose was to obtain more productive fish stocks exploitation from the lakes situated in the interior of the delta. The width of the channels varies between 9 and 18 m with a projected depth of 1.60–4.50 m. And the most difficult task was then to maintain a depth greater than 1.6 m (Rudescu et al., 1965). The natural process of sediment accumulation constantly fought, and still fights, the transformation of the Delta into an easily navigable fishing pond. The 'tapping into' the riches of the Delta relied on the previous definition of the deltaic space as a no man's land or pure wilderness. In other words, there were no other competing claims to the richness of the territory worth taking into account. In this manner, the first phase of exploration and its supposition of empty land make

possible the next stages, which see it as a matter of course that the Delta is there to be exploited by those that define it (and not, for example, by those that live there). As Gătejel (2018) chronicles, *"the engineers and commissioners in the Danube Delta quite often discarded previous local practices of river management as irrelevant for their work, instead relying exclusively on their own methods of observation and calculation"*. The engineering intervention then further solidifies the claim of the central power to the territory.

After 1878 (the Treaty of Berlin), the national borders underwent some major changes in Eastern Europe and, consequently, the Dobrogea region became part of Romania. Therefore, the new provinces needed to be connected with the older ones from a cartographical point of view, and this was a challenge.

Between 1909 and 1911, Gheorghe Vidrașcu developed a cartographical masterpiece, the most complete map of the Danube Delta at the time. The space between the arms, missing from previous materials, revealed a network of lakes and channels. Many fisheries were depicted in the interior of the Delta with permanent and temporary settlements that evoked a human presence rooted in the natural environment. The main activities were conducted along the Sulina channel, but some land-based routes could be identified along the natural levees. Those roads corresponded to animal farming activities, which increased after 1878. This revelation of the network of channels and lakes corresponds in time with the engineering modification of the main navigable branch, Sulina. Though Vidrașcu's map shows settlements and roads, so ways of using the territory by local inhabitants, the simplification of navigation that obtains in the same historical period makes it so the interior of the Delta becomes ever more accessible, and therefore ever more exploitable. This in turn gives further incentives for ignoring what Vidrașcu's map clearly shows, namely that the territory was already extensively used by local inhabitants. It is of course not for their benefit that channels are dug and dikes created (they, after all, could navigate the natural Delta just fine).

Despite the substantive interventions above, the most dramatic transformation of the Danube Delta environments occurred after WWII, when its economic function completely changed the entire ecosystem. New artificial channels were built in order to exploit the fish resources existing in deltaic lakes, to harvest reeds, or connect the fluvial arms: Roșuleț-Împutița, Magearu, Eracle-Batacu, Ciotic, Buhaz, Gotca, Iacob-Batacu, Crișan-Caraorman and Mila 35. The construction of such a dense network of canals almost tripled the water discharge toward the interior of the delta plain (Giosan et al., 2013). Between 1985 and 1990, the meanders of the Sf. Gheorghe arm were shortened, with a decrease in length from 108.2 km to 69.7 km (Fig. 3B). The consequences were considerable: an increase in the water flow speed on the artificial courses and a decrease on the natural arms with the formation of new islets.

Harvesting reed in the 1960s generated another impulse for new artificial channels. Fish farms were established over the next decade (1970–1980), some of them in the former reed enclosures (Bălteni, Maliuc, Rusca, Obretin), but many others as new projects (Chilia Veche, Stipoc, Popina, Holbina I and II, Periteașca, Perișor, Ceamurlia). It was an unsuccessful idea because, very soon, the fish farms deteriorated and had secondary effects on soil composition, salinization being the most common one. The aquaculture operations were not a big success either, due to oversized ponds, inefficient management and poor infrastructure (Van Assche et al., 2012). Agricultural polders created in the 1980s are another substantial phase of the anthropogenic impact in the Delta: Pardina, Sireasa, Murighiol-Dunavăț, Babina, Cernovca (Fig. 5). Draining and new irrigation channels transformed the deltaic space with dramatic consequences for the next years. A comparison of the maps from the beginning of the 20th century with the current distribution of water reveals a reduction in the aquatic surface by 93.2 km², mainly as a result of draining the lakes for agricultural polders. This environmental change of transforming the natural lakes and marshes into agricultural fields had complex consequences for the entire ecosystem (Fig. 5).

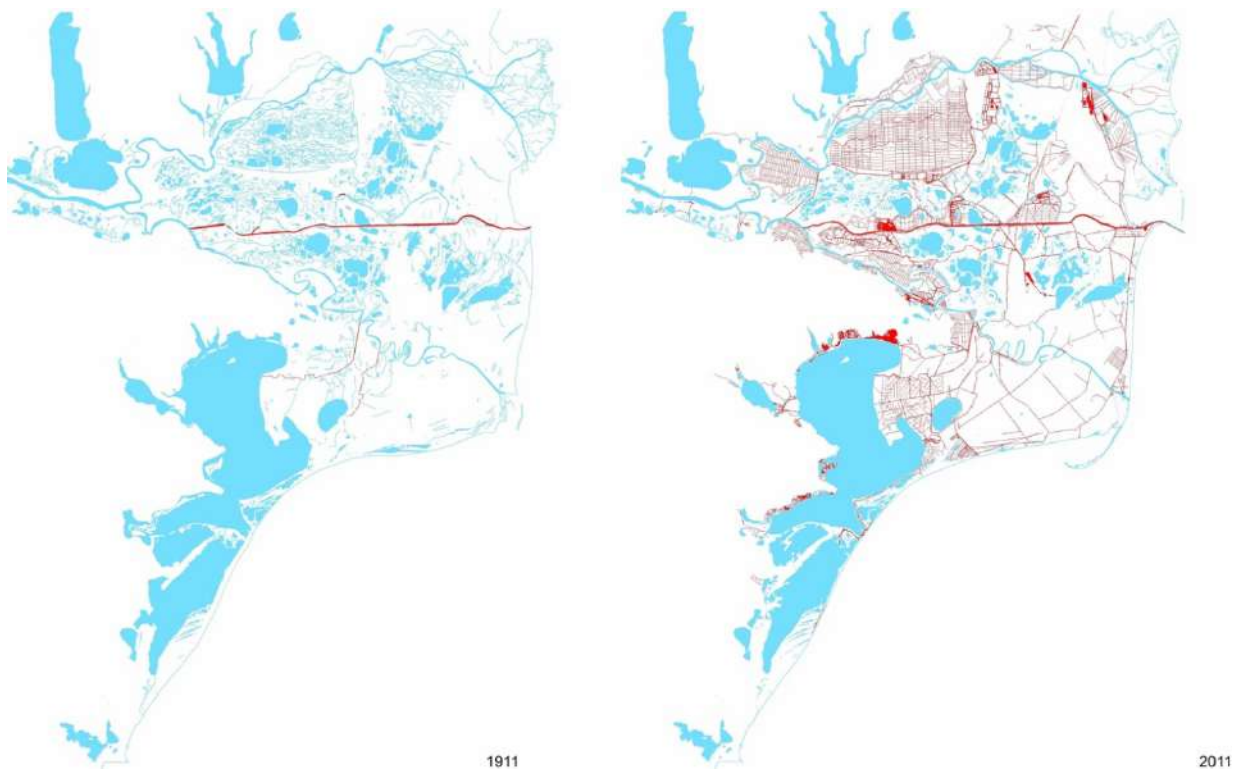


Figure 4 Anthropogenic channels cut in Danube Delta are represented with red colour in 1911 (left) and in 2011 (right)

The complex of channels, together with the shortening of the main branches, can be seen as a simplification of an otherwise labyrinthine space. Besides aiding in the exploitation of natural resources, these measures also reinforce the centre-periphery dynamic that dictates deltaic exploration. The simplification of the natural form of the Delta is done for the benefit of the representatives of the centre (of power), for whom the complex natural Delta is illegible. For the historic settlements in the Delta, however, the network of natural channels and lakes is just part of a home range that one learns to navigate from birth. As Steinberg and Peters (2015) theorize, "*territory – a political technology that combines control of land and terrain with ideas about its capacity for organisation through calculative rationality – is achieved through the control of volumes*". The deltaic environment in its natural form is best understood as a world of volume, water being the dominant element that dictates the rhythm of the land as much as that of all beings living within the delta. Its literal transformation into a territory of straight lines leading to predictable and fixed spaces

is a profound exercise of power aimed at controlling it in all of its aspects.

The planted forest is another human-induced landscape in the Danube Delta, with forestry being practised in the Rusca, Carasuhă, Pardina, Murighiol, and Sf. Gheorghe enclosures. Some species, like the Canadian poplar, have been used as an alien competitive solution to the willow and resulted in a change of the native fauna composition and of the entire ecosystem (Găstescu and Știucă, 2008). At present, the entire deltaic space is completely changed after the anthropogenic intervention (Fig. 5), and the most dramatic phase belongs to the Communist decades (1950–1989).

4.3 Aerial perspective

Topographic maps were the most frequent kind of approach in the 20th century, supplemented after 1960s by the aerial images used primarily to update the previous products. A first satellite image of the Danube Delta appears in the 1970s, but only few scientists had access to the Landsat archives, dominated in Romania by secrecy and military control. This status will be changed in the 1990s when Land-

sat and Spot images become a common tool for studying the deltaic landscape. In the present, the new Sentinel products (freely available for download from the European Spacing Agency) constitute the main resource available for a large public.

After 1991, the Danube Delta became a Biosphere Reserve, with new areas defined by this status: strictly protected areas – 9%; buffer zones – 20.7%; ecological reconstruction areas – 2.7% which have a tendency to increase over time. The agricultural polders and the fish farms still exist (7% each), but their efficiency is questioned and their degradation became the main issue. Furthermore, the statistical data associated with their distribution is very often unreliable, and it should be regarded only as an approximation.

The technological advances leading to the aerial perspective – satellite images available with increasing frequency and accuracy – makes possible a view of the territory as ecologically unified, which plays into the current ecological protection model for the area. Relying heavily on the contemporary prestige of the photographic image, the aerial view of the Delta purports to show the territory as it really is, therefore hiding more successfully than ever the collusion between cartographic knowledge and the exercise of power. The contemporary phase of deltaic cartographic knowledge further obscures the multiple dimensions of local knowledge for whom the Delta is a place to be inhabited and not a picture to be divided into zones (see Figure 5).

The phase of ecological protectionism in the history of the Delta should not be understood as a hiatus of anthropogenic impact, but rather itself as the next phase of impact. The Danube Delta Biosphere Reserve regulates the use of the territory strictly and therefore decides which changes are acceptable and which are not. Even when an area is strictly protected, its characteristics change precisely because of the ways in which it is cut off from the greater deltaic space. Fencing forests off (as in the case of Letea), regulating where semi-feral cattle can go, where and what can be fished or hunted, all in the context of a changing climate, is best understood as the next phase of human-induced changes. The civilizing assumption that the Delta is fundamentally empty makes the progression from phase

(1) to (4) inevitably into a progression from empty wilderness to resource exploitation. Even though locals were supposed to exist, their presence was always understood as benign, that is to say as one with no fundamental agency or power over their fate or that of the natural environment. This is contradicted by practices such as reed burning or tree planting that significantly alter the local environment in non-harmful ways. The current protectionist phase fits perfectly well within an anthropogenic history that privileges the rights of the outsider, of the one with the power to describe.

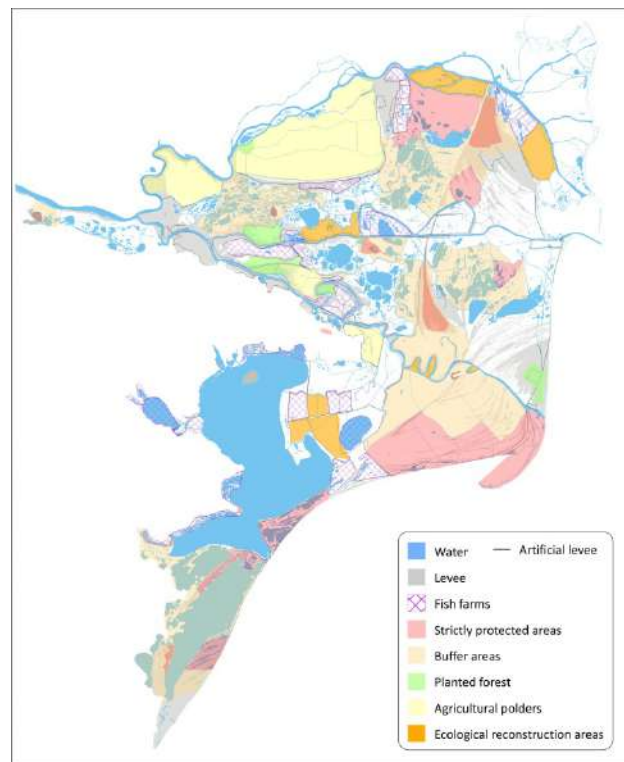


Figure 5 Different types of human-induced environments in the Danube delta (data from Gâstescu and Știucă, 2008)

The engineering of the Delta has further reinforced its accessibility, the volumetric character of the natural environment having been considerably flattened. The current phase of ecological protectionism has not significantly restored the Delta to pre-engineering states and, if current trends are any indication, it will not do so in the foreseeable future.

As with the previous periods of cartographic knowledge aiding certain uses of the territory, so the ecological protectionism prevalent today fa-

vours certain activities over others. The power to determine which areas are strictly protected and which are available for local economic activities rests with the central authorities, with no local consultations being carried out. This is in keeping with the history of deltaic exploration we have described. The territory as a nature reserve is of great use to tourist operators and big external investors. This is itself in keeping with the current economic expansion of financial capital into areas previously off-limits to its power. The dominant forces in the area are the authorities that impose the rules and regulations of the nature reserve and the economic investors that run business in the area, increasingly focused on leisure activities. The Delta as nature reserve is a story that sells, and the current powers are as invested as naval powers once were in keeping – and showing, through maps – the Delta as a natural wonder.

5. Conclusions

In the last decades, efforts have been made to grasp the evolution of the Danube Delta landscape (Antipa, 1914, 1942; Brătescu, 1922; Banu and Rudescu et al., 1965; Panin, 1989, Giosan et al., 2006, Vespremeanu-Stroe et al., 2017). Every delta has its particular issues, and it is difficult to propose a general solution due to the differences in the climate regimes, the liquid and solid discharges, the political administrations or the anthropogenic pressures of each case. Even though each delta is unique, the approach advanced in this study can prove useful for understanding historical changes elsewhere.

We showed that the first stage of cartographic knowledge involved a descriptive and pictorial style with confusing information and many errors related to locations, but those maps were not supposed to aid access to the heart of the Delta but rather to represent it as part of the known world and to inscribe it in the religiously dominated symbolic world order of the day. The mathematical cartography development corresponded to an increase in anthropogenic interventions, the map both aiding and being aided by new means of exploration. The military approach involved bathymetric measurements for fluvial navigation and accurate charting of the coast. The search for economic benefits in the Dan-

ube Delta began in the second half of the 19th century, but for other deltas, this could be the case of earlier or later explorations. The natural resources exploitation, which is still ongoing today, produced some dramatic transformations in the deltaic environment: initiation of artificial channels, reed harvesting, fish farms, agricultural polders or even industrial hot spots that completely changed the environment. The political and ideological changes of the early 1990s led to a new status, in this case the Biosphere Reserve one, with its new functional zones overlapping with the traditional ones.

Fishing and animal husbandry were the first human activities that started to transform the Delta, later followed by transit navigation and commercial activities. Economic exploitation led to the most significant changes in the deltaic environment through agriculture, the creation of fish farms, forestry enclosures, and even industrial activities. Even though the habitats' conservation umbrella is protecting some small parts of the Delta, new ecological restoration areas will still have to change the functionality of unsuccessful agricultural, aquaculture or forestry experiments, in the difficult context of climate change. And in spite of all these human interventions in the natural environment, the Danube Delta still remains a global biodiversity hotspot and a better-preserved wetland, compared to other similar deltas (Giosan et al., 2014). Learning about the inaccessible deltaic spaces is a complicated process, which requires a long period, spent first along the coast, then along the main branches and finally inland. Therefore, a cartographical diagnosis could be the most suitable tool for understanding the complex changes occurring in river deltas and how these are inseparable from wider political and ideological considerations. This approach could be complemented with local accounts of the territory in order to reveal the cultural meanings of land-use practices, without which any sustainable use of the deltaic territory is doomed. The phases of deltaic exploration have all supposed an empty land, and this has both made possible the exploitation of the territory by outsiders and has buried land-use practices that could be used in future adaptive strategies. The coastal lowlands that are located less than a meter above sea level will be inundated by the

turn of the century (Giosan et al., 2014), therefore raising new challenges. In order to be prepared for such scenarios, we need to turn to the past to explain our surroundings. An appropriate solution would be to re-assess our historical maps and to examine them for a better understanding of our future. We can also re-evaluate the assumptions at the basis of our cartographic relation with the Delta. There, we stand a chance to discover historic local practices that can show the way to a more benign anthropic impact and a fairer distribution of power. The protectionist phase, in incorporating human agency, might then be able to propose a dynamic and adaptable human ecology, and not a return to a past that never existed.

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Developing a neuro–fuzzy system to classify drainage sub-basins according to erosion processes on the Island of Lefkas, Greece

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ABSTRACT

In this paper we attempt to classify drainage sub-basins according to their erosion risk. We have adopted a multi-step procedure to face this problem. The input variables were introduced into a GIS – platform. These variables were the vulnerability of the surface rocks to erosion, topographic variations, vegetation cover, land use and drainage basin characteristics. We constructed a fuzzy inference mechanism to pre-process the input variables. Next we used neural-network technology to process the input variables. The system was trained to 'learn' and classify the input data. The output of this procedure was a classification of the sub-drainage basins related to their risk of erosion. This neuro–fuzzy system was applied to the island of Lefkas (Greece).

KEYWORDS

erosion; fuzzy sets; neural networks; GIS; Ionian Sea

1. Introduction

Environmental settings such as geological, geomorphological and climatic conditions affect drainage structures. Characteristics of rocks and soils, geographical distribution and intensity of rainfall, along with micro-topography, determine surface and

underground water. Runoff can be either Hortonian or not depending on the infiltration capacity of surface rocks. Emmet (1978) has stated that the flow characteristics of the surface waters may produce rills or sheet flow, and as a result, weathered material can be transferred by water movement. Tem-

poral and spatial aspects also affect erosional processes. Splash erosion, caused by raindrop impact, seems to greatly influence the detachment of material. According to Kirkby (1978), the rate of soil erosion is proportional to the distance from the divide multiplied by the power of slope gradient. The phenomenon of soil erosion has been studied by many researchers (Stocking, Elwell, 1973; Elwell, 1978;

Wischmeier and Smith, 1978; Morgan et al., 1984; Nearing et al., 1989; Briggs and Giordano, 1992; Thornes et al., 1996; Baturst et al., 1996; Koulouri and Giourga, 2007; Cerdan et al., 2010). Some methodologies of artificial intelligence, especially fuzzy set theory, have also been previously applied to erosion issues by Burrough (1989), Burrough et al. (1992), Gournelos et al. (2005).

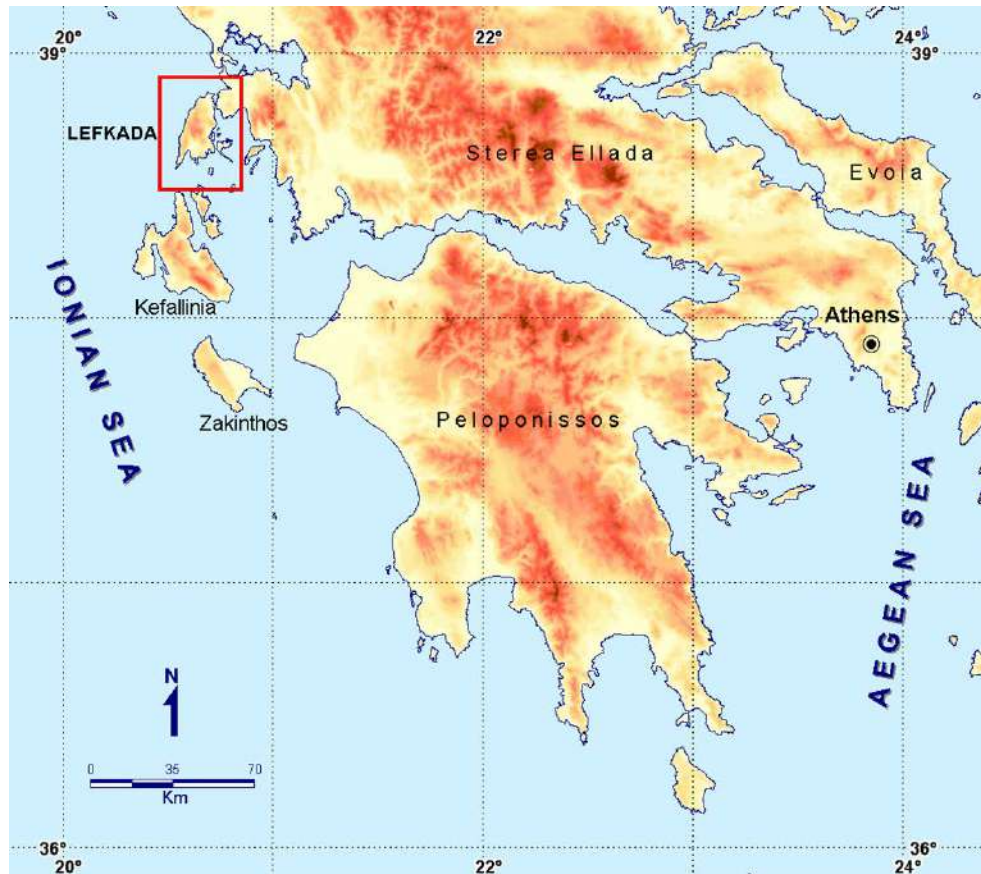


Figure 1 Location of the study area

In addition to the natural modification of drainage basins, human induced factors also play an important role: activities such as the expansion of agricultural areas, building activity and fires greatly affect erosion rates. Soils are key to ecosystems and human societies, and their critical importance requires a better understanding of how they evolve through time. However, identifying the role of natural climate change versus human activity (e.g. agriculture) on soil evolution is difficult (Rothacker et al., 2018).

The Mediterranean region is particularly vulnerable to erosion. This is primarily owed to major

land-use changes during the last decades and the long dry periods followed by intensive and erosive rainfall, particularly on areas of steep slopes with fragile soils, (Van der Knijff et al., 1999).

In this paper, the combination of fuzzy set theory and neural network technology constitutes an innovative methodology that was used to study erosion processes, by generating classifications of autonomous drainage units. All variables affecting and defining erosion were considered here, and the methodology was applied on the island of Lefkas.

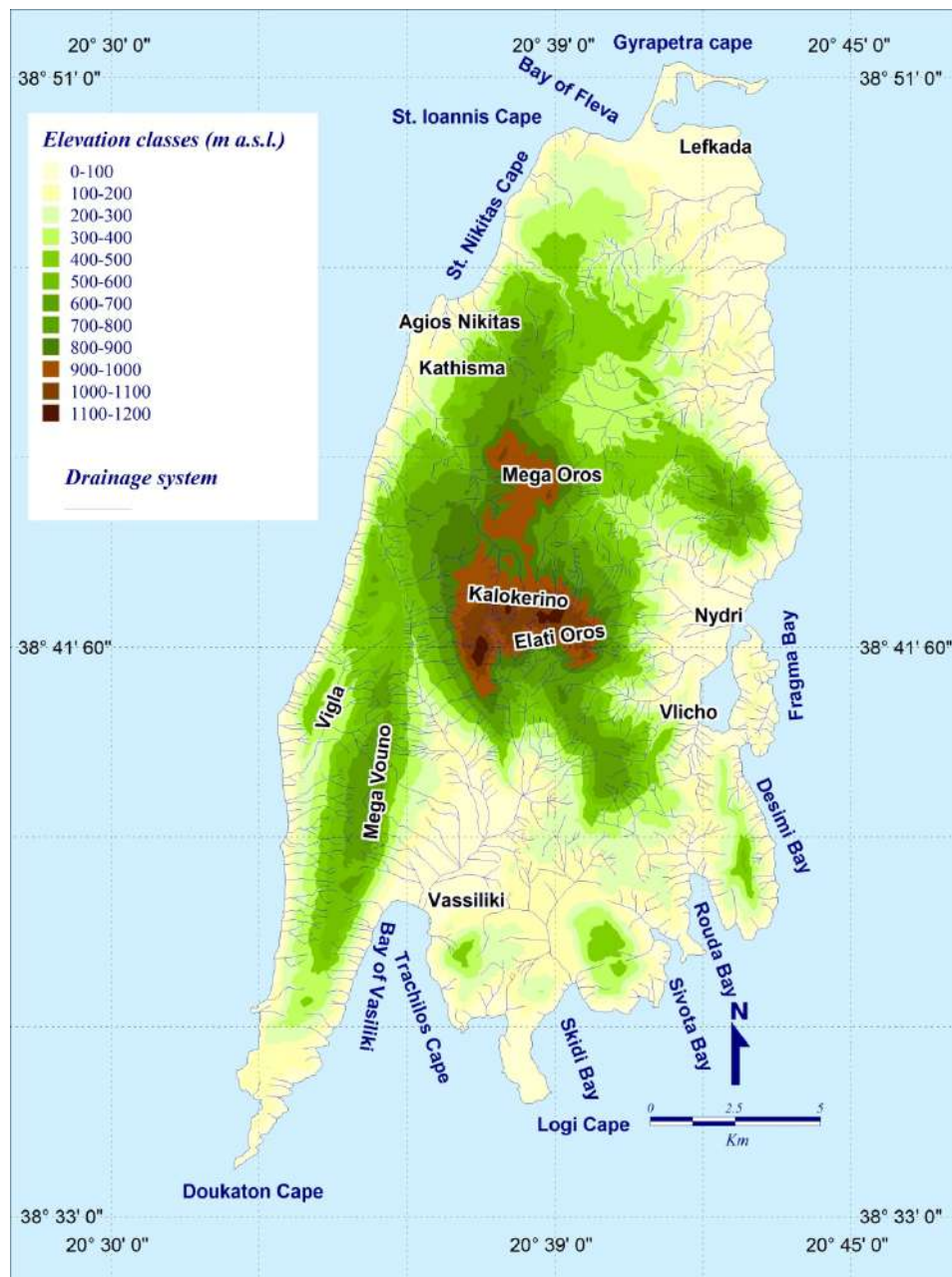


Figure 2 The hipsometry of Lefkas island

2. Study area

Lefkas Island is located in the Ionian Sea, in the western part of Greece (Fig. 1) and covers an area of 294 km². The island is mainly characterized by mountainous and hilly landscapes and a few basins of low altitude (Fig. 2). It is composed of alpine and post-alpine rock formations, some of which are extremely vulnerable to erosion processes. The geological study of Lefkas has been the object of several studies, such as Bornovas (1964), Verginis

(1976), Livaditis and Verikiou-Papaspiridakou (1987) and Underhill (1989). The study area is composed of two geotectonic units, Ionian and Paxoi. The greater part of Lefkas is covered by the Ionian unit (210 km²) while the Paxoi unit is located in the southwest part of the island (covering an area of 63 km²). The main lithologies of the alpine substratum are limestones and flysch. There are also some Neogene and Quaternary formations in the internal and coastal basins of the island (Fig. 3). It becomes clear

from the lithological composition that some of the island's surface rock formations are extremely vulnerable to weathering and erosion processes.

The island's annual precipitation levels are relatively high (mean value: 942.3 mm/year. for the

years 1975–1997; Nastos, 2001). Lefkas belongs to a zone of high seismicity. A recent earthquake with a magnitude of 6.4 (14/08/03) resulted in a number of landslides in its western part, where steep slopes are dominant (Fig. 2).

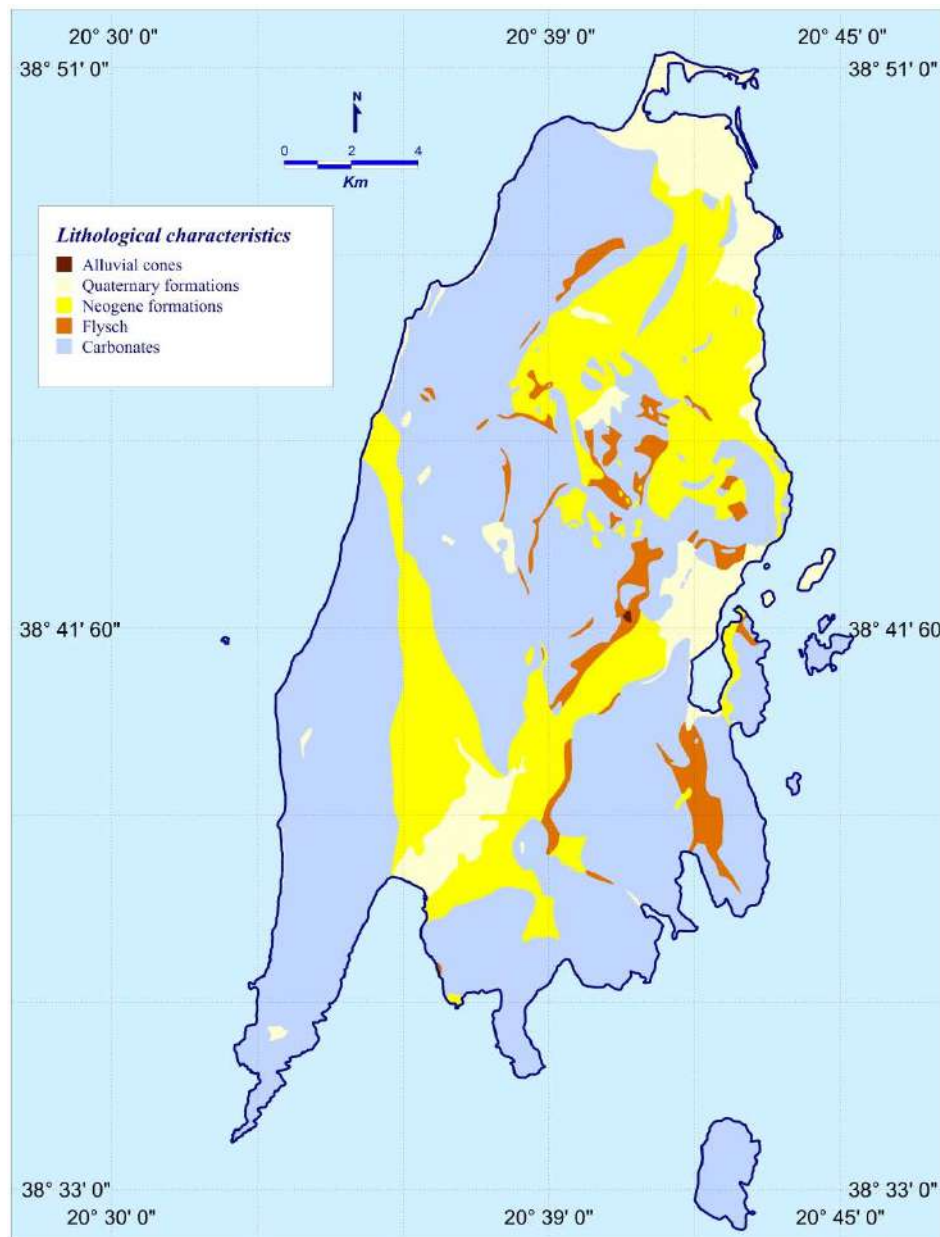


Figure 3 Lithological formations of Lefkas island

3. Methodology

In order to study a process as complex as erosion, it was deemed necessary to develop a GIS-based procedure. Firstly, data from various thematic maps, aerial-photos and fieldwork were digitized and im-

ported into a GIS (MapInfo 2006). Secondly, the definition of the input and output variables was accomplished. The input variables that were processed are the following: vulnerability of surface rocks to erosion, vegetation cover and land use, drainage characteristics and slope gradient.

The three first variables are closely related to the erodibility of rocks. Rock vulnerability to erosion is the result of several physical and chemical factors, such as rock composition and structure and the presence or absence of protective vegetation. Vegetation and land use firstly control the infiltration mechanism and hence water runoff (sheet flow or rills), and secondly they affect the impact of precip-

itation (Kirkby, 1995). The drainage network is the result of an established surface and subsurface flow system. The drainage system characteristics were studied by using the drainage density variable. The slope gradient determines the surface flow regime. This flow system becomes very erosive when the terrain morphology is dominated by steep slopes.

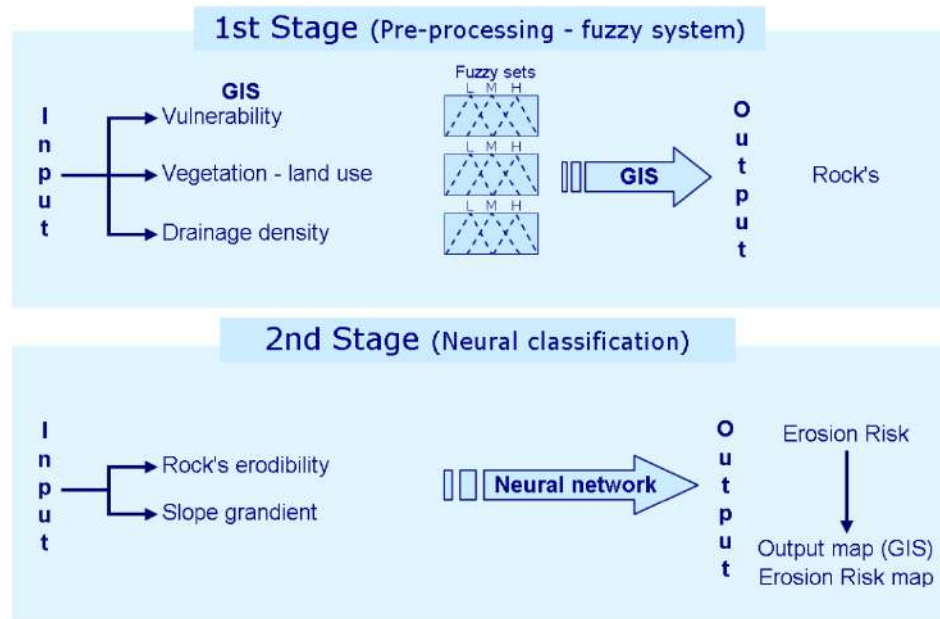


Figure 4 A neural-fuzzy system to produce erosion risk map

Erosion risk was the output variable. The sub-basins of the drainage system constituted the basic units for the evaluation of the input variables.

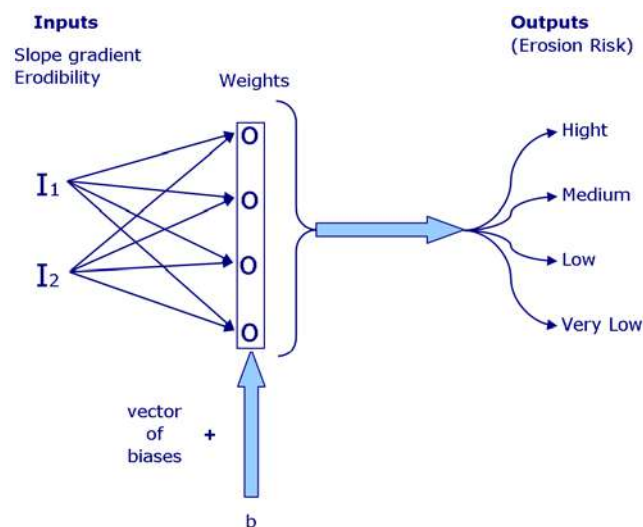
The next step was to establish an interference mechanism with which to transform the input parameters into output parameters. This process consisted of two stages: a fuzzy-based transformation and the creation of a neural network system (Fig. 4). In the first stage the input variables used were: a) rock vulnerability to erosion, b) vegetation – land use, and c) drainage density. The output variable was the rock's erodibility to erosion. All the aforementioned variables present problems of imprecision in terms of their quantification and the definition of their spatial boundaries. For this reason, it was necessary to use fuzzy set theory (Zimmermann, 1991; Klir and Yuan, 1995) (Fig. 4). The fuzzy model was produced by assigning an appropriate membership function to all variables involved and by using logical rules as a transformational mechanism

(Mamdani and Assilion, 1975). The formulation of fuzzy logical rules (Table 1) was based on numerous field observations and the authors' experience of erosion processes. The system's output is of a fuzzy nature. Therefore, defuzzification was carried out, using the centroid technique.

In the second stage we developed an artificial neural network model to classify the drainage sub-basins. In this model the input variables were rock erodibility, as derived from the fuzzy model, and slope steepness. In the neural network model, we constructed a three-neuron layer to classify the sub-basins according to the degree of erosion risk (Fig. 5). We used competitive learning architecture, which is based on the distance between data input and the weight. The winning neuron is the one with the minimum distance. This process is repeated until all input data are classified (Kohonen, 1982; Rumelhart and Zipser, 1985).

Table 1 Fuzzy interfaces rules

If VULNERABILITY IS High	&	VEGETATION – LAND USE INDEX IS Low & SLOPE IS High–Medium	Then ERODIBILITY of ROCKS Is High
If VULNERABILITY IS High	&	VEGETATION – LAND USE INDEX IS Low & SLOPE IS Low	Then ERODIBILITY of ROCKS Is Medium
If VULNERABILITY IS High	&	VEGETATION – LAND USE INDEX IS Low–Medium & SLOPE IS Low	Then ERODIBILITY of ROCKS Is Medium
If VULNERABILITY IS Medium	&	VEGETATION – LAND USE INDEX IS Low & SLOPE IS High	Then ERODIBILITY of ROCKS Is High
If VULNERABILITY IS Medium	&	VEGETATION – LAND USE INDEX IS Low & SLOPE IS Medium–Low	Then ERODIBILITY of ROCKS Is Medium
If VULNERABILITY IS Medium	&	VEGETATION – LAND USE INDEX IS Medium & SLOPE IS High	Then ERODIBILITY of ROCKS Is Medium
If VULNERABILITY IS Medium	&	VEGETATION – LAND USE INDEX IS Medium & SLOPE IS Medium	& Drainage density IS Medium Low Then ERODIBILITY of ROCKS Is Low
If VULNERABILITY IS Medium	&	VEGETATION – LAND USE INDEX IS Medium & SLOPE IS Low	& Drainage density IS Medium–Low Then ERODIBILITY of ROCKS Is Medium

**Figure 5** A three–neuron layer was used with biases

The erosion risk was the output variable of this final stage. The implementation of the aforementioned procedures was achieved using a G.I.S. (MapInfo 2006) and the Matlab software platform (Mat-lab 2005). The fuzzy model and the artificial neural network model were constructed in a Matlab environment loosely coupled with the G.I.S. The values of the input variables were obtained by digitizing topographic and geological maps (1:50.000 Geographic Military Service; Bornovas, 1964), by analyzing aerial photos (1:33.000, 1986) and satellite images (Landsat, 1980) and through field observations.

**Figure 6** The drainage basins of the island of Lefkas

In the fuzzy model, triangular functions were adopted in the pre-processing stage. All variables were characterized by fuzzy set values, and were expressed with the appropriate membership function. In the present study, “low”, “medium” and “high” define the different degrees of rock vulnerability. Rock vulnerability values for Lefkas were

based on empirical knowledge, field observations and published material (Kuenen, 1956; Leopold et al., 1964; Sparks, 1965; Bolton, 1979; Selby, 1987). We assigned high vulnerability values to the Quaternary and Neogene formations (marls, conglomerates), medium values to the flysch formations and low values to the limestones. The slope gradient and

the drainage density variables were obtained from the topographic maps, through GIS analysis. The vegetation and land use variable was based on aerial photos, satellite images and field data. The island's vegetation cover consists of trees, bushes and agricultural areas. Land use has largely changed as a result of tourist developments over the last decades.

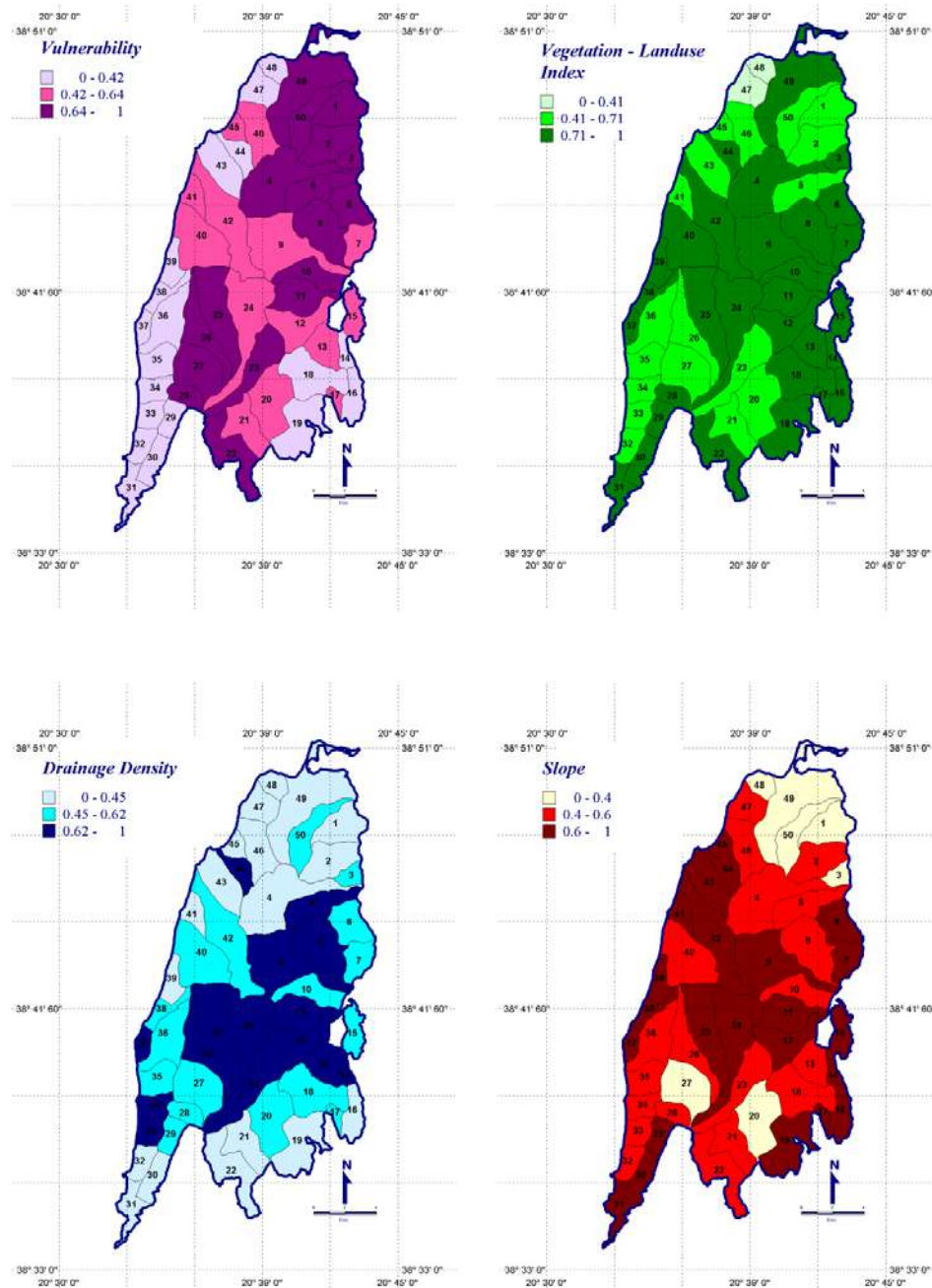


Figure 7 The input variables used in order to extract the erosion risk variable, a) vulnerability, b) vegetation–land use, c) drainage density and d) slope

Table 2 Input variables

Id	Vulnerability Normalized	Vegetation – Land use Index Normalized	Drainage Density Normalized	Slope gradient Normalized
1	1.00	0.62	0.36	0.26
2	0.92	0.54	0.44	0.42
3	0.98	0.85	0.59	0.39
4	0.74	0.90	0.27	0.49
5	0.73	0.64	0.68	0.43
6	0.79	0.81	0.48	0.63
7	0.61	0.85	0.48	0.68
8	0.67	0.94	0.71	0.51
9	0.52	0.98	0.62	0.66
10	0.75	0.85	0.58	0.52
11	0.77	0.90	0.73	0.83
12	0.56	0.97	0.62	0.80
13	0.53	0.90	0.69	0.59
14	0.39	0.95	0.66	0.72
15	0.53	0.83	0.60	0.72
16	0.38	0.97	0.45	0.94
17	0.47	0.88	0.46	0.85
18	0.42	0.91	0.61	0.49
19	0.37	0.80	0.37	0.62
20	0.49	0.64	0.46	0.34
21	0.52	0.62	0.39	0.41
22	0.79	0.98	0.26	0.55
23	0.85	0.58	0.68	0.60
24	0.58	0.96	0.79	0.65
25	0.70	0.90	0.78	0.72
26	0.82	0.55	1.00	0.43
27	0.81	0.64	0.46	0.37
28	0.64	0.91	0.48	0.49
29	0.36	0.99	0.54	0.81
30	0.37	0.98	0.31	0.65
31	0.39	1.00	0.14	0.74
32	0.36	0.61	0.30	0.55
33	0.38	0.52	0.69	0.60
34	0.38	0.52	0.67	0.55
35	0.36	0.64	0.61	0.57
36	0.37	0.61	0.59	0.44
37	0.36	0.95	0.70	0.87
38	0.36	0.82	0.58	0.72
39	0.36	0.81	0.41	1.00
40	0.50	0.90	0.50	0.60
41	0.55	0.57	0.24	0.61
42	0.44	0.98	0.46	0.61
43	0.39	0.60	0.19	0.63
44	0.39	0.83	0.62	0.80
45	0.42	0.57	0.32	0.75
46	0.48	0.62	0.34	0.58
47	0.38	0.11	0.18	0.44
48	0.36	0.16	0.14	0.40
49	0.81	0.81	0.24	0.27
50	0.79	0.59	0.56	0.37

We used the same fuzzy classes for the drainage density and vegetation – land use variables. Thus, low index values for vegetation and land use indicate drainage basins with sparse or no vegetation, medium values indicate underdeveloped vegetation cover, whereas high values signify relatively dense vegetation cover.

The drainage network is defined by topography and surface rock formations. We have identified 50 drainage sub-basins for the whole island (Fig. 6). In order to combine data layers, all the original data were normalized by dividing them by their maxi-

um value. All normalized values are presented in Table 2.

Using the rules of Table 1, mapping of the three input variables was accomplished, rock vulnerability, vegetation/land use and drainage density (Fig. 7a, b, c), in order to create the output variable, the rock's erodibility. The erodibility variable was then combined with slope steepness (Fig. 7d) within the developed neural network to create the erosion risk variable. The final output variable of this neuro-fuzzy system was the erosion risk for all the drainage sub-basins on Lefkas (Fig. 8).

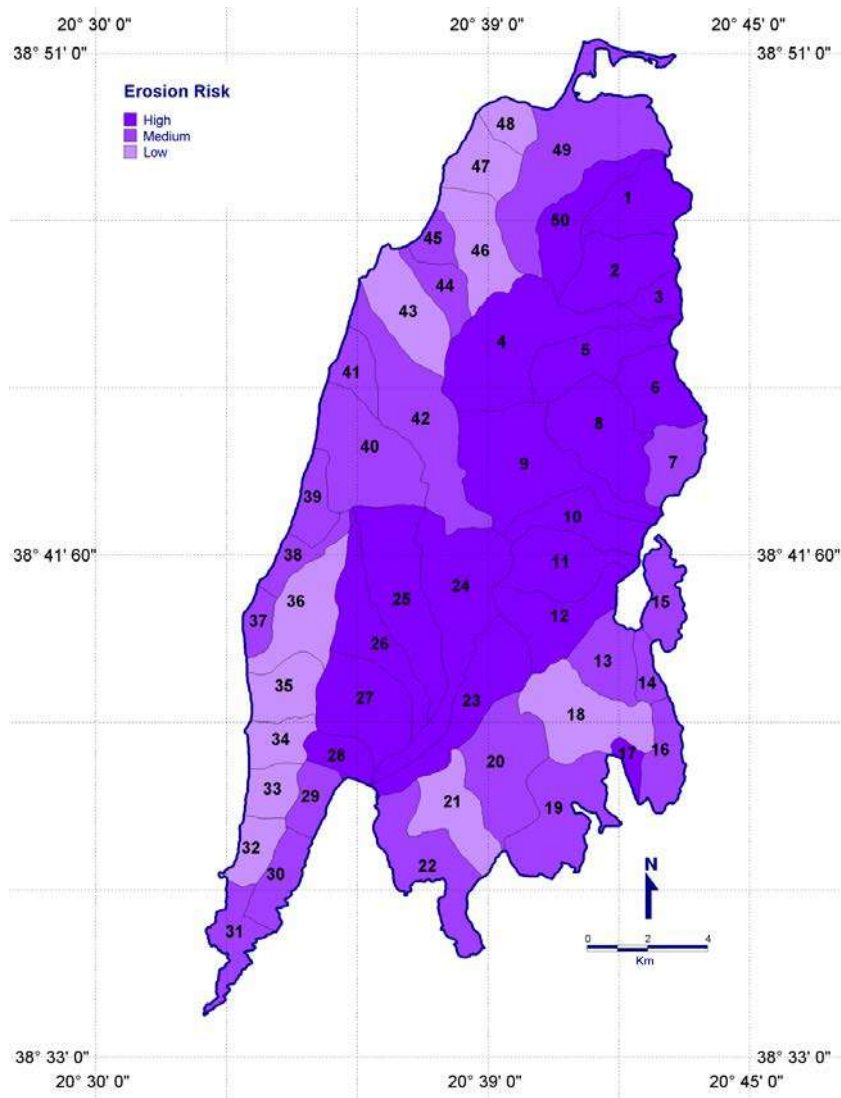


Figure 8 Erosion Risk index of all the drainage sub-basins on Lefkas

4. Discussion and conclusions

In the case of Lefkas island, we established three classes of erosion risk: high, medium and low. A

sensitivity analysis was carried out in order to test the model's behavior to minimal changes in the input variables. As was expected, small perturba-

tions in the values of the three input variables cause no significant changes in the output variables (i.e. the erosion risk index).

We have established three classes of erosion risk: high, medium and low (Fig. 8). The drainage sub-basins belonging to the upper classes (high erosion risk) are situated in the northeastern and the southern parts of the island. The medium risk drainage sub-basins are situated in the central part, while the low risk basins are located mainly in the eastern part. The spatial distribution of the erosion risk index variable may be primarily related to the spatial distribution of the lithological formations: where highly vulnerable rocks are present (Neogene and Quaternary) high risk drainage sub-basins are also located, whereas low risk drainage sub-basins are situated in locations with rocks that are more resistant to erosion (limestones). The whole classification scheme takes into account all the input variables involved, especially the nature of surface rocks and topography.

This neuro-fuzzy approach for studying erosional processes is an extremely practical tool because of its ability to handle and classify qualitative data. Erosion risk maps, such as the one produced for Lefkas Island, may be used to identify areas of risk and for the implementation of mitigation measures to protect agricultural and urban areas. This relatively simple model may be used in local- and region-level planning, environmental protection and hazard assessments. This methodology may be further improved by using smaller scale grid-based data as elementary input variables.

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Recent landslides from Iași Metropolitan Area

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ABSTRACT

The creation and analysis of landslide inventories is the basic approach for starting the landslide hazard to risk analysis chain analysis. In order to perform this kind of analysis for the Iași Municipality (North–eastern Romania) we created based on aerial imagery and LiDAR data a landslide inventory of the landslide events that happened in the last 100 years. In total, we identified and delineated 518 landslide events: 51.5% translational slides, 19.7% slumps (rotational slides), 17% flowslides and 11.8% flows. The majority (74%) of the landslides events happened between 1956 and 1984, those before 1956 representing 16%, while those triggered after 1984 representing 10%. This situation reflects the very humid period between 1960 and 1990, when the majority of the landslide events were triggered. Beside the typology we have also identified the triggering factors and we can conclude that rain-storms coupled with previous rainy years were the main triggering factors, especially for Holocene old and relict landslides in areas with land use and anthropic disturbances created by the built up area expansion. The analysis of the landslide area frequency density is showing that the shape of the distribution curve is similar with what was reported in the literature, which is signaling that this inventory can be used further in landslide susceptibility modelling and validation. Further analysis of these events can also pinpoint scenarios for vulnerability and risk analysis.

KEYWORDS

recent landslides; Iași Municipality; North–eastern Romania

1. Introduction

The “landslide problem” (Brabb, 1991) require the study of landslides for the reduction of landslide risk. Landslide hazard is defined as the probability of landslide occurrence within a certain area and in a

certain period (Varnes et al., 1984; Guzzetti et al., 1999; Guzzetti, 2006). A landslide inventory is a database (Carrara and Merenda, 1976; Guzzetti, 2006; Guzzetti et al., 2012) which describe in depth landslide events, their spatial and temporal location, and

other attributes (their typology, causes and their effects). Landslide inventories represent the base information needed for landslide susceptibility modelling and validation (Carrara et al., 2003; Guzzetti, 2006) and for vulnerability scenario assessment (Alexander, 2005), crucial elements in the landslide hazard to risk chain (Reid, 1998; Crozier and Glade, 2005; van Westen et al., 2006).

We present a spatial database of landslides events from Iași Metropolitan Area, which happened during the last 100 years and which can be used for landslide susceptibility modelling and validation. For these events, the historical data, remote sensing data and the field evidences are enough to provide the full picture of the landslide morphology, the typology, the activity and to allow precise delineation. We argue that this database and its possible extension to include more events is of crucial importance for the landslide hazard and risk modelling for Iași County.

1.1 Study area

Iași Metropolitan Area (IMA) is located at the contact between the Central Moldavian Plateau and the Jijia Hills (Fig. 1). Beside Iași Municipality, IMA contains the territories of 18 communes (1092 km²): Movileni, Popricani, Victoria, Rediu, Aroneanu, Lețcani, Valea Lupului, Holboca, Ungheni, Miroslava, Tomești, Țuțora, Mogoșești, Ciurea, Bârnova, Schitu Duca, Comarna and Prisăceni.

In the Jijia Hills the monoclinic structure and the homoclinal shifting of the river network (Jijia and Bahlui) influenced the development of a repetitive pattern of valleys and ridges (Fig. 2), with cuesta scarp slopes and cuesta dip slopes (Băcăuanu, 1968; Ioniță, 2000; Niculiță, 2011). The lithology (Fig. 3) is predominantly composed from siltstones and mudstones with sandy intercalations (Martiniuc et al., 1956; Brânzilă, 1999; Ionesi et al., 2005; Dill et al., 2011) from the Bessarabian Cryptomactra Clays Formation (under 125–150 m a.s.l.) and from mudstones and siltstones intercalated with sands from the Bessarabian Bârnova–Muntele Formation (230 m thick appearing over 125–150 m a.s.l.). The cuesta hills have a caprock of fluvial (2 to 5 m thick) and

loess (5 to 25 m thick) deposits (Martiniuc and Băcăuanu, 1959; Martiniuc and Băcăuanu, 1966; Schram et al., 1977) and their absolute altitude do not exceed 220 m a.s.l. (Rediu Hill, Aroneanu Hill, Miroslava Hill, Galata Hill).

At the contact between the Jijia Hills and Central Moldavian Plateau the caprock structure generated an impressive escarpment: Iași Cuesta Escarpment, with an amplitude of up to 350 m (Băcăuanu et al., 1980; Ungureanu, 1993). Here in the Central Moldavian Plateau over the Bârnova–Muntele Formation there is a caprock of calcarenites, oolitic calcarenites and quartz arenites intercalated with sands and mudstones (Jeanrenaud and Saraiman, 1995; Brânzilă, 1999; Ionesi et al., 2005; Dill et al., 2011) which end the Bessarabian suite, the Repedea Formation (30 m thickness, with lateral facies variations toward south and east which show an increase in sands and sandstones and a decrease of limestones – Jeanrenaud and Saraiman, 1995; Ionesi et al., 2005).

Over the Bessarabian rocks, a 100 m transgressive thick Kersonian suite appear on top of the hills, which is composed mainly from sands (Ionesi et al., 2005). Toward south over the Kersonian suite there are the Meotian rocks from the Nuțasca–Ruseni Formation (Fig. 4). This formation is around 160 to 260 m thick and it is composed from sands and cineritic sandstones (10–80 m) and sands and mudstones (150–180 m) (Jeanrenaud and Saraiman, 1995; Ionesi et al., 2005). Because of the resistant caprock the hills from this area have relatively flat and elongated ridges, while the hillslopes have a relief of up to 250 m.

The superficial deposits are various (Martiniuc and Băcăuanu, 1959; Cernătescu et al., 1966; Băcăuanu et al., 1980; Barbu et al., 1987): (i) deluvial deposits and altered bedrock on steep hillslopes where landslides are not present, (ii) mass movement deposits where landslide are present, (iii) fluvial terrace deposit (sands with basal gravels) covered by loess on cuesta dip slopes from Jijia Hills, (iv) soils or eluvial deposits on ridges and plateaus, in the Central Moldavian Plateau, (v) clayey and silty alluvial deposits on floodplains.

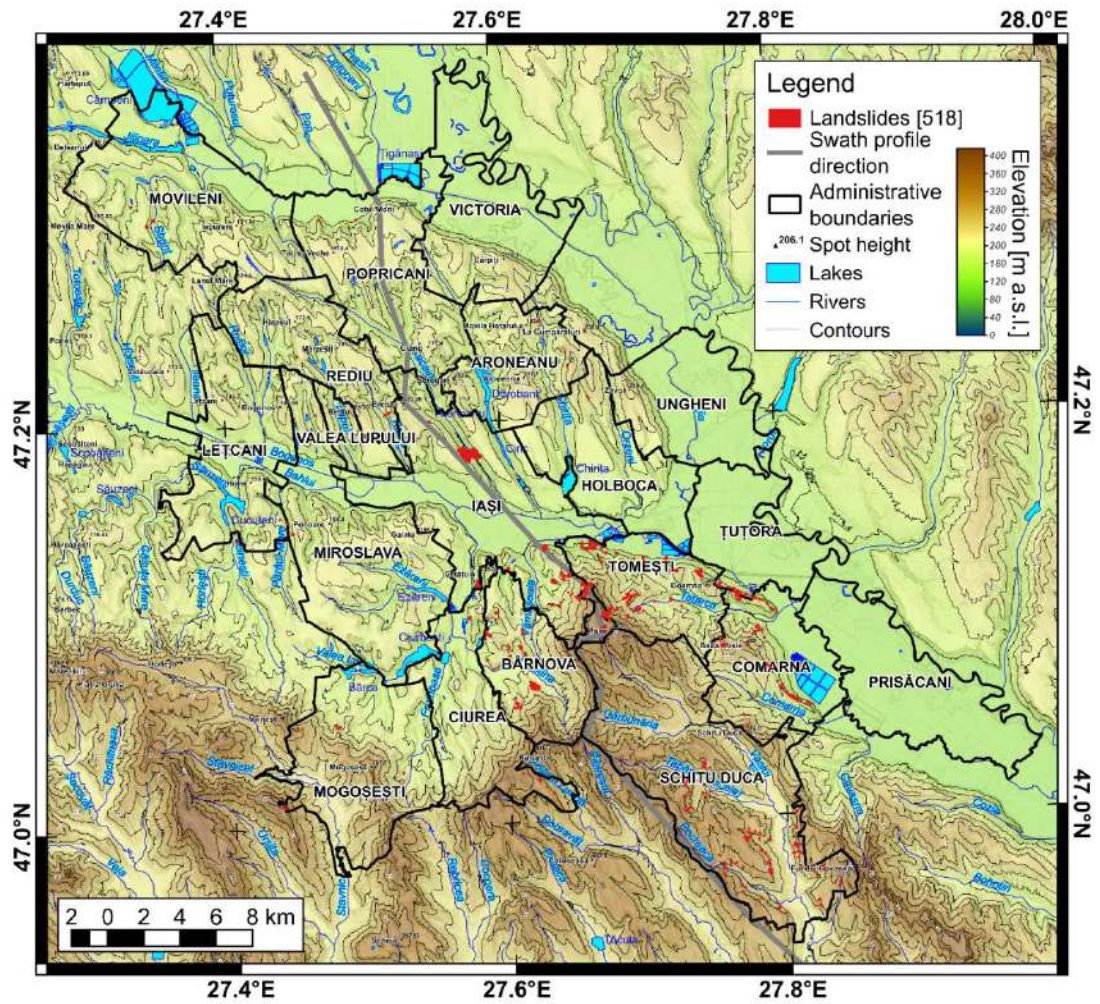


Figure 1 Geographic position of Iași Metropolitan Area

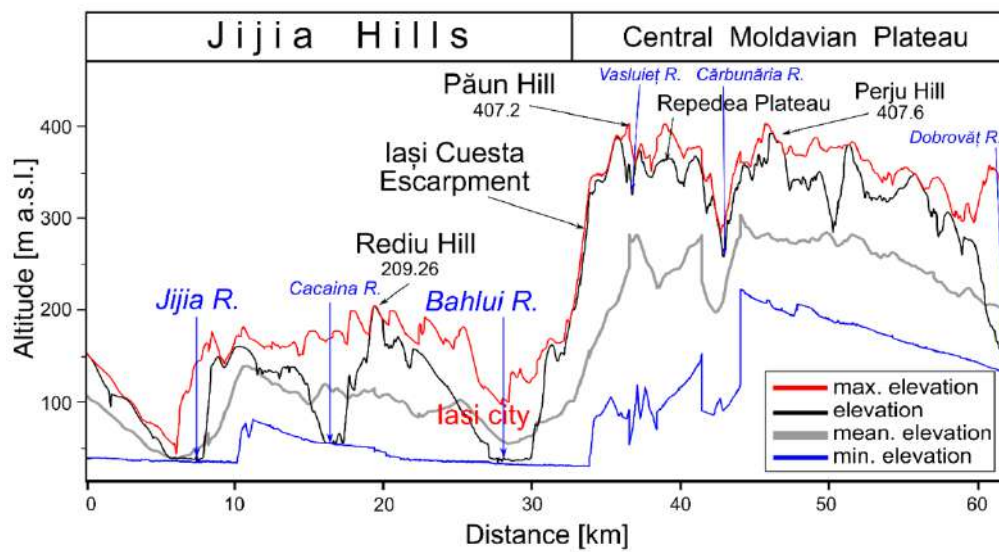


Figure 2 Swath profile (15 km wide) through the study area (the vertical scale is exaggerated in order to allow the recognition of all the elevation profiles)

1.2 Landslide history

North–Eastern Romanian hilly area is well known for its relict and old landslides developed especially on cuesta scarp slopes (Niculiță and Mărgărint, 2014; Niculiță et al., 2016; Mărgărint and Niculiță, 2017), generating slope deposits which are very susceptible to landslide reactivations (Silion, 1965). Recent landslides which still keep the landslide morphologic elements under the current dry climatic conditions happened after Medieval times (Niculiță et al., 2016). Iași surroundings were covered by forests in the medieval times (Giurescu, 1976), especially toward south (Iași–Bârnova forest) and south–west (Căpoteștilor forest), but also toward east (Bîc forest), south–east (Bohotin forest) or north (Copou and Cîric forests – Tufescu, 1932) which were cleared in the modern period as the city extended from the Palatul Culturii fluvial terrace (Băcăuanu and Martiniuc, 1966) toward the surroundings (Tufescu, 1932). In this context we believe that the reactivations of the old and relict Holocene landslide (Martiniuc and Băcăuanu, 1982) bodies and scarps (Silion, 1965), while happening probably also during medieval times, became more frequent after, when the forest clearance and human pressure intensified ("Galata din Vale"/"Galata de Jos" Monastery which was affected by landslides between 1579 and 1582 – Grigoraș, 1943; Cârciuleanu, 1991; Ureche, 2017).

Areas with reactivations are known for Iași surroundings in the following locations: the Copou Hill north–eastern hillslope (1911, 1932–1933, 1934, 1940–1942, 1961, 1969–1970, 1984, 2017), the Păcurari neighborhood, the Șorogari Hill western hillslope (1969–1970), the Galata Hill northern hillslope (1932–1933, 1941–1942, 1960, 1970, 1971, 1973, 1974–1975, 1979), the Cârlig Hill western hillslope (1981), for Cetățuia (1979–1980) and the Bucium area (1973) as reported by various authors (Macarovici, 1942; Silion, 1965; Băcăuanu, 1970; Palade and Băcăuanu, 1971; Brișcan, 1980; Martiniuc and Băcăuanu 1982; Schram et al., 1977). In recent times landslide reactivations were reported in Todirel village (5 March 1999), Munteni Neighborhood, Grădina Botanică, Șapte Oameni, Moara de Vânt. Beside these areas, where the landslide mechanism have a strong natural component, there

are several places where the landslide mechanism is highly influenced by anthropic processes, especially along roads: on communal road DC49 toward Hadâmbu Monastery (2000 to 2017), Bahlui river bank (2015), on county road DJ248D in Bucium (2000–2010) and along the national road DN24 between Păun and Poieni (2005).

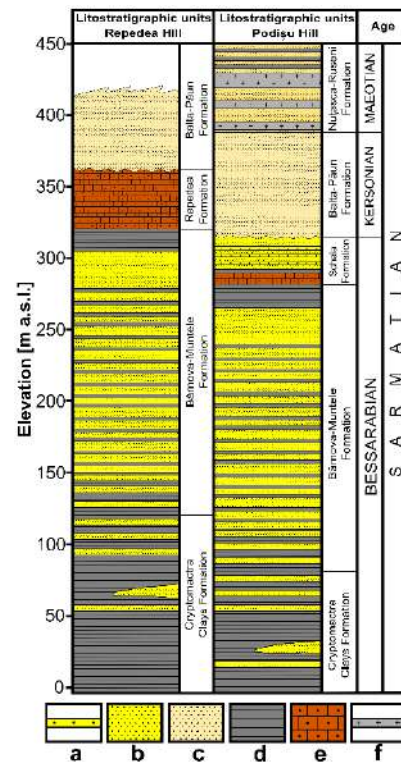


Figure 3 Lithostratigraphic columns of IMA geology (Jeanrenaud, 1971, Jeanrenaud and Saraiman, 1995; Brânzilă, 1998, 1999; Ionesi et al., 2005): a – sandstones, b – sands, c – deltaic sands, d – siltstones, e – oolitic limestones, f – cinerites

A geomorphological landslide inventory for Iași municipality was prepared by Necula and Niculiță (2017) and a historical inventory for Iași County was prepared by Niculiță et al. (2017). The slow moving displacement rates were identified for the 2014–2017 period for the north–eastern hillslope of Copou Hill by Necula et al. (2017).

2. Materials and methods

2.1 Materials

Historical map sources (1:20000, 1:5000, 1:2000 topographic maps from the 1920–1960 period) were used to identify landslides which are older than

1956 and LiDAR was used for the delineation of these landslides. Especially for the Copou Hillslope LiDAR data had to be interpreted also based on the historical reports.

The availability of four series of aerial images (1956–1964, 1971–1984, 2003–2005, 2008) and a 0.5 m spatial resolution LiDAR DEM acquired in 2012, allowed us to identify landslide events that happened between 1956 and 2015.

2.2 Methods

Landslide events were mapped as polygons only when all the landslide elements were identified on the aerial imagery or LiDAR (Ardizzone et al., 2007; Petschko et al., 2015; Niculiță et al., 2016): crown and scarp, flanks, toe, rough landslide body (Fig. 5).

In unforested areas, these elements were quite easy to be depicted. In forested areas the field work and the use of LiDAR point clouds helped us to understand the morphology of the landslides. The typology was assessed using the Hungr et al. (2014) extension of the Cruden and Varnes (1996) classification. We identified the following types of landslides: (i) recent soil/earth planar/translational slides (type 12) develop mainly on steep hillslopes previously affected by old or relict landslides; (ii) on old landslide scarps, rotational soil/earth slumps are frequent (type 11); (iii) old landslide deposits with a mixture of clays, silts and sands or the sandy bedrock are affected by flowslides and earthflows (type 20 and 26).

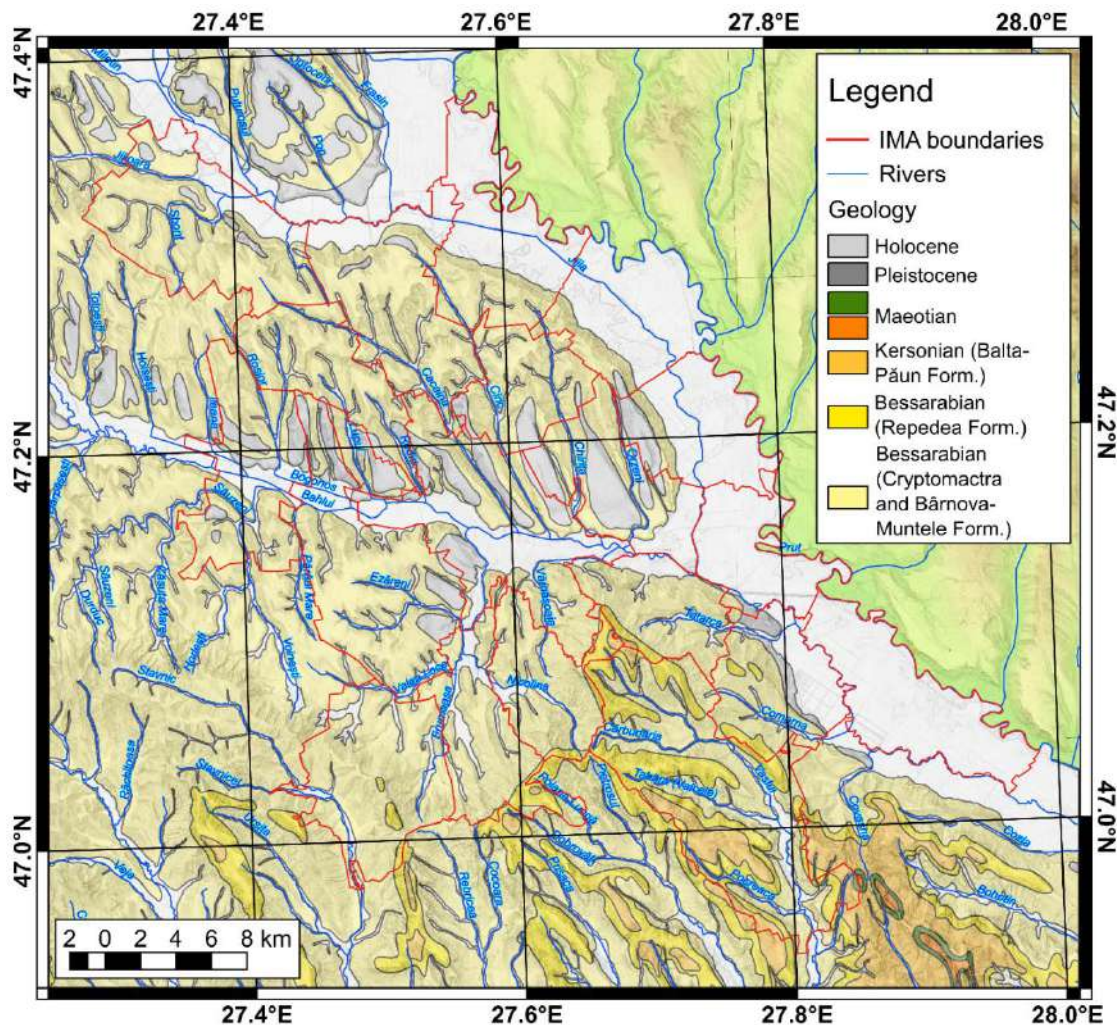


Figure 4 Geologic map of IMA (compiled after the 1:200 000 Geological Map of Romania, Saulea et al., 1966, completed with maps and information from Jeanrenaud and Saraiman, 1995, Brânzilă, 1998, 1999, Ionesi et al., 2005)

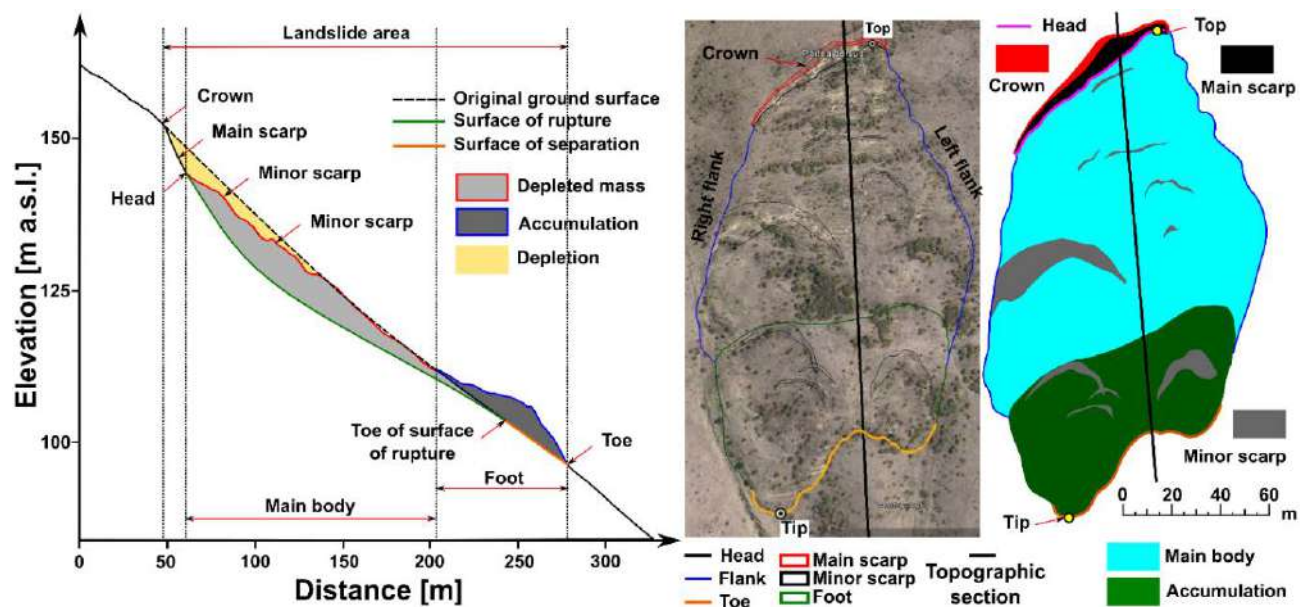


Figure 5 Landslide delineation elements for the post 1984 landslide localized on the northern hillslope of Miroslava Hill (the elements were considered regarding the Cruden and Varnes, 1996 work) on a topographic section (left), aerial imagery (center)

3. Results and discussions

We have identified and mapped 518 landslide events (Fig. 1). The majority of the identified landslides are translational slides (51.5 %). Slumps (rotational slides) and flowslides are the other type of landslides considering the frequency (19.7 % and 17%). Flows represent 11.8%. The association of the process type with the displaced material is very hard to be performed because the slope deposits are very similar with the geological deposits.

The majority of the events appear on previous slided areas, as reactivations of the scarp or of the main body. Translational landslides are more frequent on old and relict landslide bodies, but flows are also present. Slumps (rotational failures) and flowslides are the most frequent on old and reactivated landslides scarps. Flows are also present along scarps. Some of the landslides that were triggered before or after 1964 and before 1984, were levelled after 1970 by the antierosional measures taken in the study area, so do not appear on the LiDAR dataset (Fig. 6 – 4th row).

The magnitude of the identified landslides is low (under 140 000 m²), under what is considered in the literature magnitude 1 (Malamud et al., 2004), their length doesn't exceeding 700 m and with widths in general under 50–100 m (Fig. 7). The

shape of the area frequency density is similar with that of landslide inventories triggered by specific events, situation which allows us to consider this landslide inventory to be representative for the present day conditions in the study area (Fig. 7) and to be usable in landslide susceptibility modelling and validation. The low magnitude and the fact that almost all the identified landslides happened outside populated areas show that landslides do not represent a real day by day threat for the population of IMA, but the situation could change in the future, considering the extension of the built up area (Iațu and Eva, 2016; Doru, 2018) and the climate changes.

There are some exceptions, several areas being known for frequent reactivations: Copou north-east hillslope (Țicău and Sărărie area), Păcurari, Aurora, Șipoțel, Brândușa, Galata, Cetățuia, Cărlig, Todirel. The several notable exceptions generated population displacement during the last 100 years and still represent a real threat for the houses that remained. Considering that IMA is in continuous expansion of built area, the landslide susceptibility modelling (Necula and Niculiță, 2017) should be taken into account for planning purposes, since new built area will cover territories where these small magnitude landslides are frequent.

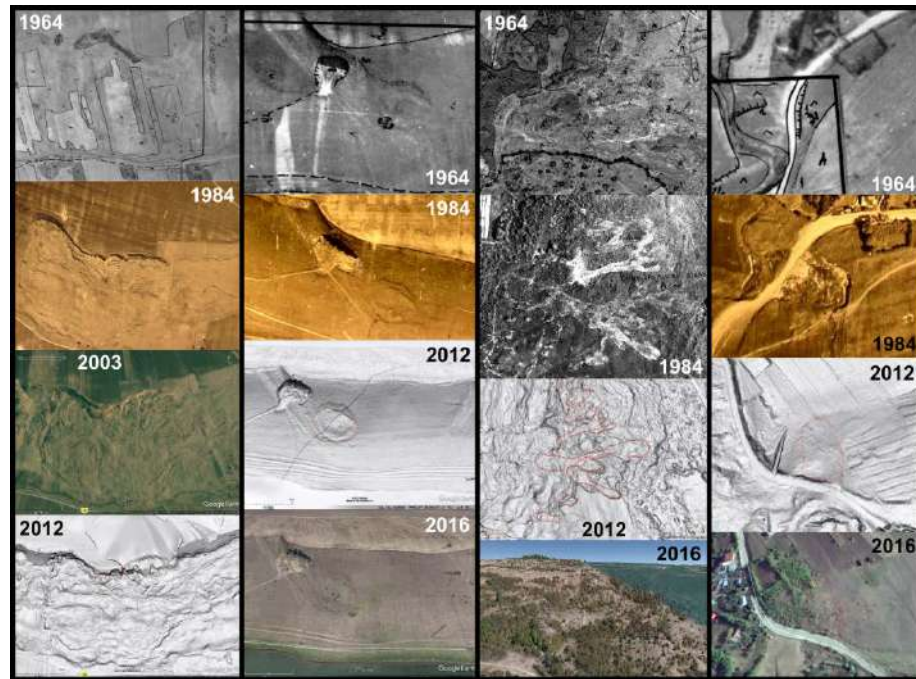


Figure 6 Landslide typology on various resources used for identification and mapping: slumps north of Rusenii Noi triggered after 1964 (1st row from left), translational slide triggered before 1964 (2nd row from left), flows and flowslides from Bucium hillslope triggered after 1964 (3rd row from left), translational slide south of Orzeni triggered after 1964 (4th row from left)

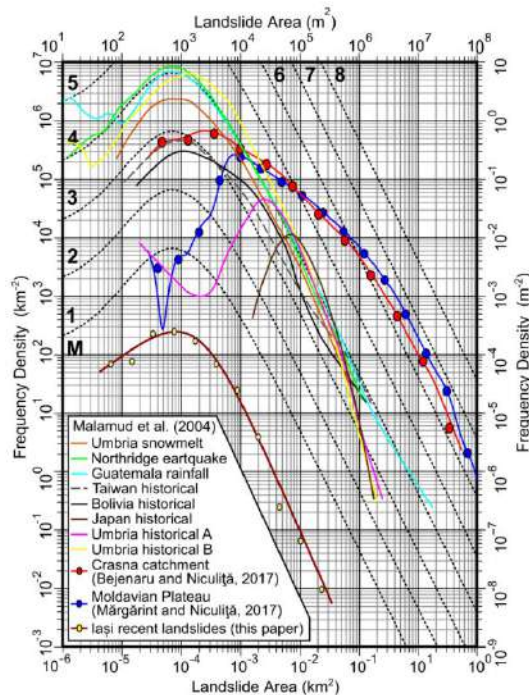


Figure 7 Landslide area frequency density for the produced landslide inventory and comparison with other inventories from the Moldavian Plateau (Bejenaru and Niculiță, 2017; Mărgărint and Niculiță, 2017) and around the world (Malamud et al., 2004)

A strong temporal pattern exist, the majority of the landslides appearing between 1956 and 1984 (74%). The landslides which were triggered before 1956 but probably not before 1890–1920 (because the disturbances are not present of the topographic maps) represent 16%, while those triggered after 1984 represent 10%. The 1960–1990 period is known to be a wet period, compared with the period after 1990, which is considered a dry period (Fig. 8 and 9). Future rainy period might increase the frequency of landslides in areas which today are considered stable.

The precondition factors of the recent landslides from Iași Metropolitan Area are the monoclinic and caprock morphostructure, the lithology (intercalations of clays and sands), the climate and the land use (Mărgărint and Niculiță, 2017). The mean annual rainfall is increasing from Iași (102 m a.s.l. – over 550 mm) to Bârnova (395 m a.s.l. – over 770 mm), mainly because of orographic precipitation induced by the western air masses movement over the Central Moldavian Plateau (Roe, 2005; Mihăilă, 2006; Minea, 2012; Pelin, 2015). In the southern part of the study area the forest is dominant, while in the northern

part of the study area the agricultural fields and the pastures are dominant. In the central part, in Iași Municipality there is a high density of built up areas (Doru, 2018).

The preparatory factors are the pre-existence of old and relict landslides (Niculiță et al., 2016; Mărgărint and Niculiță, 2017), the rainfall variation and the land use changes. Rainfall variation in the study area was mainly related to the increase of the rainfall quantity in May–July compared to the August–October, in 1891–1920, 1920–1935, 1940–1953, but especially after 1960, until 1991 (Pelin, 2015). The most frequent duration of consecutive years with rainfall excess is 2 years (Mihăilă, 2006). In the last two centuries the anthropic pressure increased, through forest clearance, agricultural fields, orchard, vineyard and settlement extension (Doru, 2018). The rainfall variability and the changes

in land use affected especially the stability of the Holocene landslides scarps, where at the contact between the permeable fluvial terrace and loess deposits with the impermeable siltstones and mudstones important aquifers develop (Macarovici, 1942; Silion, 1965; Palade and Băcăuanu, 1971; Brișcan, 1980; Martiniuc and Băcăuanu 1982; Schram et al., 1977). The saturation of the materials from the base of the scarps and of the landslide body deposits generated the majority of the reactivations, and also controls the landslide type. Landslides under forests are in general of small magnitude and disturb the forest only in the scarp area, although especially in areas where anthropic disturbances appear important landslides were triggered (Horticulture Faculty event – Palade and Băcăuanu, 1971). Areas which are recently deforested very often supported landslide reactivations.

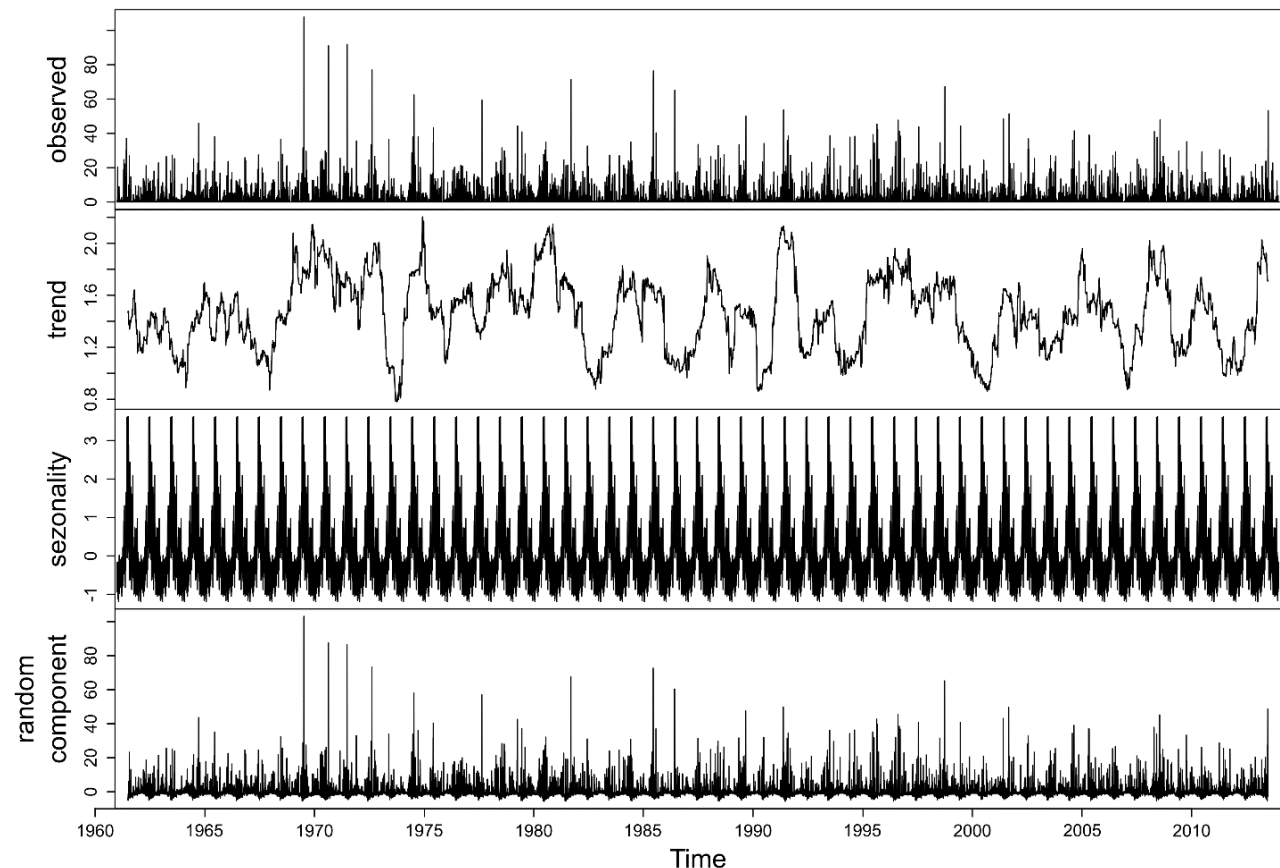


Figure 8 Additive decomposition of daily rainfall temporal data for Iași meteorological station, according to ROCADA dataset (Dumitrescu and Bîrsan, 2015)

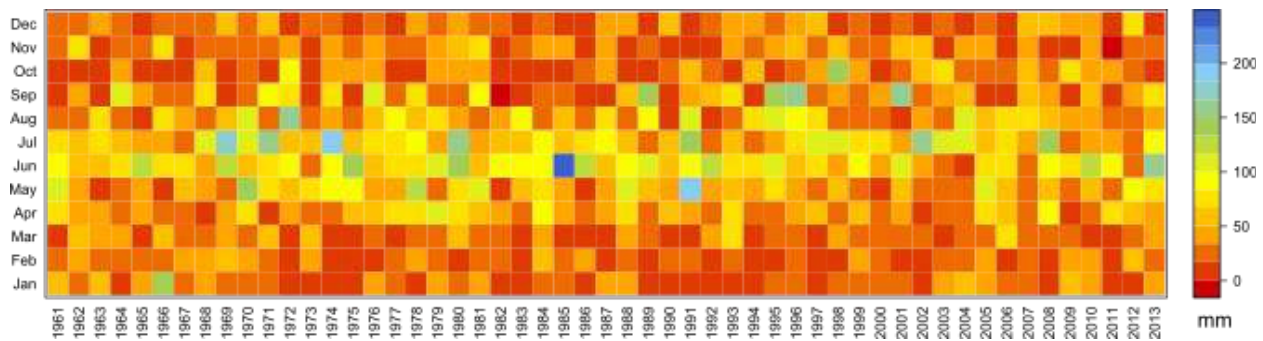


Figure 9 Mean monthly rainfall data for Iași Municipality, according to ROCADA dataset (Dumitrescu and Bîrsan, 2015)

4. Conclusions

We present a database of 518 landslides events from Iași County, which happened in the last 100 years. For these events, the historical data, remote sensing data and the field evidences are enough to provide the full picture of the landslide typology and morphology, of the landslide conditional, preparatory and triggering factors, of landslide magnitude and temporal evolution, of immediate effects on humans and infrastructure and the implications on future human activity in the area.

Among the studied events the Copou hillslope, Iași city, is one of the most interesting case of landslide reactivations during the last 100 years, with more possible future reactivation events, and which generated the biggest damages, fortunately without human life loss. While some of the presented landslides affected human settlements and infrastructure, the majority are cases where the landslides happened outside settlements and infrastructure, but the extensions of the built up area inside the metropolitan borders is to be expected. Also, in the case of an increase in the amount of rainfall, the frequency of landslide events might also increase.

Landslide inventories which describe in depth landslide events, their spatial and temporal location, their causes and their effects are very important for landslide susceptibility modelling and validation and for vulnerability scenario assessment. This database and its possible extension to include more events is of crucial importance for the landslide hazard and risk modelling for Iași Metropolitan Area

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Permafrost investigations in Iezer Mountains, Southern Carpathians

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ABSTRACT

This paper assesses the possibility of permafrost occurrence in Iezer Mountains using field observations, GIS analysis and thermal investigations in the field. Rock glaciers as the main mountain permafrost footprint in the landscape are mapped using cross validation from two independent inventories, analysed and classified in terms of size, altitude, morphology, air temperature and vegetation cover. Measurements of spring temperatures, bottom temperature of snow (BTS) and continuous (near) ground surface temperature (GST) were applied to check for permafrost presence. Autumn temperature of 20 alpine springs are analysed and grouped by their source, temperature and apparent discharges. BTS was applied on 3 sites from the upper part of Colților Valley in 2012 and 2018 along GST in Tambura scree in 2012–2018 period. 16 rock glaciers are considered to exist in Iezer Massif, a much smaller number than previously assumed. We argue that several other landforms are rock–ice features and should be inventoried and analysed in corresponding with minimal past permafrost creep. Rock glaciers are poorly developed and have a faded morphology because of low debris availability, short periods of time with favourable climates and less supportive mountain morphometry. Alpine springs indicate possible permafrost occurrence in one rock glacier and one talus slope and BTS reveal sporadic patches of permafrost down to 1900 m asl in Colților Valley. GST supports the multiannual stability of permafrost indicating BTS on a multiannual scale and suggests the great importance of autumn–early winter temperatures and snow interplay for ground cooling and permafrost maintenance. We argue that 11 rock glaciers from Iezer Mountains are probably inactive. MAGST of 1.1 – 3.2 °C suggest a strong thermal gradient in the blocky layer if permafrost is truly present. Alpine screes possibly underlain by permafrost seem not to be disturbed by any modern climate changes.

KEYWORDS

mountain permafrost; rock glaciers; rock–ice features; talus slopes; Iezer Mountains

1. Introduction

Permafrost is ground understood as soil or rock and included organic matter and ice content that has a temperature of 0 °C or below for at least two consecutive years (van Everdigen, 1998). It is exclusively defined on the basis of temperature regardless the texture, water content, lithology and degree of induration (Washburn, 1979, p. 21). Permafrost is not a form but a thermal state of the ground without an explicit and direct appearance in the landscape (Dobinski, 2011). This phenomenon has been intensely studied in the last decades in the context of climate change because permafrost can release a large quantity of organic carbon that can accelerate global climate change (Schuur et al., 2015). The contemporary warming and degradation of permafrost is being documented both at high latitude and altitude and it is expected to continue (Etzelmüller et al., 2011; Hipp et al., 2012; Kokelj et al., 2017; Oliva and Fritz, 2018). Warming permafrost in mountain areas can be a source of hazards inducing an increase in slope instability (Ritter et al., 2012; Deline et al., 2015; Haeberli et al., 2017). On the other hand, mountain permafrost is a valuable resource of water that communities from arid and semiarid zones especially can rely on in the context of rapid glaciers retreat (Viviroli et al., 2011; Rangecroft et al., 2016; Drewes et al., 2018).

Striking geomorphologic features created by permafrost occurrence in mountain areas are rock glaciers which form by creeping of debris accumulated by rockfalls or deposited by glaciers and containing ice of glacial or periglacial origin (Barsch, 1996; Berthling, 2011; Knight et al., 2019). However, permafrost presence does not necessarily imply rock glaciers development if other environmental preconditions are not met (debris accumulation and ice formation) and also rock glaciers existence does not necessarily express the actual permafrost occurrence. The latter is the case of relict rock glaciers (no frozen content) that results from intact rock glaciers (active and inactive, i.e. with supersaturated and moving-/thin, under saturated, stable permafrost, respectively) subjected to permafrost thawing caused by the climate transition to a warmer phase. These rock glaciers can be used as palaeoclimate

indicators (Hughes et al., 2003; Aoyama, 2005; Kellerer-Pirklbauer et al., 2012; Moran et al., 2016).

Openwork debris deposits determine negative thermal anomalies in comparison to other land cover types and consequently are the most favourable structures for permafrost development (Harris and Pedersen, 1998; Gruber and Hoelzle, 2008; Otto et al., 2012). Discontinuous permafrost zone is restricted to the lowest altitude of the active (moving by the process of creep) rock glaciers while sporadic permafrost zone imply the presence of inactive or relict rock glaciers only with small patches of permafrost (Baroni et al., 2004; Kellerer-Pirklbauer et al., 2015). Talus slopes are also favourable to permafrost development in their lower sectors because of ground air circulation induced by density differences (chimney effect) occurring at both low (Morard et al., 2010; Stiegler et al., 2014) and high altitude (Scapozza et al., 2011; Kenner et al., 2017). In order to check for the presence of permafrost and to determine its spatial extent, an entire set of indirect and direct methods of detection should be used along multiple years in order to avoid short term climate effects.

Rock glaciers occurring in Romanian Carpathians were first assessed by Ichim (1978). Regional inventories (Urdea, 1992 and Onaca et al., 2017a) indicated the ubiquitous occurrence of rock glaciers in Southern Carpathians highest massifs that were used for palaeoclimate reconstructions (Urdea, 1998). Several studies indicate the relict state of rock glaciers supported by their subdued morphology and thermoclastic depressions in Făgăraş Massif (Nedelea, 2006; p. 189; Simoni, 2007, p. 293; Andra, 2008, p. 292). Other studies suggest the rather scarcity of rock glaciers in the alpine areas of Parâng Mountains (Săndulache et al., 2015). Indirect investigations (thermal and geophysical) proved more widespread potential distribution of permafrost in granitic massifs (Retezat and Parâng) in comparison to the crystalline ones (Făgăraş) (Urdea, 1993; Szepesi, 1998–1999; Kern et al., 2004; Nedelea, 2006, p. 195; Vespremeanu-Stroe et al., 2012; Onaca et al., 2013; 2015; Popescu et al., 2015). Continuous monitoring of temperature introduced the hypothesis of probable occurrence of perennially frozen rockwalls

above 2350 m altitude in Southern Carpathians (Vasile et al., 2014; Popescu et al., 2017a).

Low altitude permafrost was also investigated in Romanian Carpathians at Detunata Goală, Apuseni Mountains (Urdea, 2000; Popescu et al., 2017b).

The objectives of the present paper are to describe and analyse the geomorphological characteristics and significance of rock glaciers and to increase the knowledge on the potential permafrost distribution in Iezer Mountains.

2. Study area

Iezer Mountains are located in the eastern part of the Southern Carpathians between Făgăraş in the west and Piatra Craiului massifs in the east (Fig. 1). They have a total plane surface of 535 km² from which almost 14 % (74 km²) are located above 1800 m asl and a maximum height of 2470.2 m in Roşu

peak. The main ridge has a NE–SW orientation with two areas of high altitudes in the SW around Iezerul Mare peak (2461.8 m) and in the NE around Păpuşa peak (2391.4 m asl). The lithology of the massif is represented mostly by crystalline schists i.e., mica schists and garnet paragneiss, two mica paragneiss with garnets, gneiss, chlorite and muscovite–chlorite schists (Codarcea and Dimitrescu, 1968; Patruşiu et al., 1968). The massif has a pronounced asymmetry with elongated interfluvies and valleys on the SE side and short and abrupt slopes on NW side. The transversal profiles of most of the valleys are asymmetric with abrupt western/southern slopes and gentle eastern/northern slopes for the north–south and east–west oriented valleys respectively. Large sectors from the main ridge and other interfluvies of Iezer Mountains are represented by the remains of Borăscu peneplain.

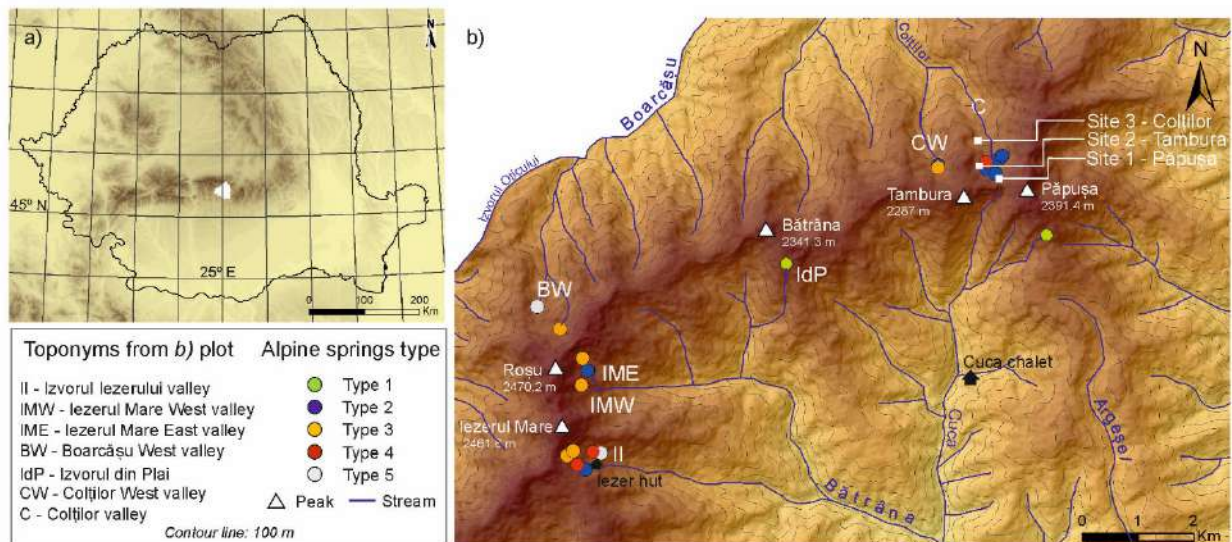


Figure 1 Study area. Location of Iezer Mountains in Southern Carpathians, Romania (a), study sites in Colţilor Valley and the distribution of alpine springs (b). Contour interval 100 m

The upper treeline does not surpass 1870/1830 m asl on the NW/SE slopes respectively as an effect of land cover (talus deposits) and anthropic intervention (deforestation) respectively, while the sub-alpine shrubs (*Pinus mugo*) climb usually up to 2100 m and exceptionally up to 2270 m asl (Săvulescu, 2014, p.122–129).

The past glacial erosion was less intense in Iezer in comparison with other massifs from Romania presenting a total of 38 cirques with a mean floor

elevation of 1880 m asl some of them clustered around plateaus (Mindrescu and Evans, 2017). The mean altitude of the end moraines was calculated to be 1395 m asl and the total surface of glaciers at their maximum extent 60 km² (Urdea and Reuther, 2009). From a structural point of view, the northern and north–western cirques are obsequent, the eastern ones are subsequent while the southern cirques are consequent (Szepesi, 2007, p.56). Rockwalls occur mostly on the western or southern slopes of

the valleys and they present a high degree of erosion. Talus deposits occur at the base of the rock-walls and also on the less inclined slopes. Large and compact areas of debris deposits can be found in the glacial cirques around the two highest areas of the massif of Păpușa and Roșu – Iezerul Mare peaks, respectively. The current active geomorphologic processes are solifluction, debris flows, snow avalanches and rock falls.

3. Methodology

3.1 Geomorphological mapping of rock glaciers

We mapped rock glacier landforms using orthophotos (produced by National Agency for Cadastre and Land Registration, 2012) and aerial images from Google Earth and field research. In mapping rock glaciers, we looked for their typical morphology signs like longitudinal ridges in the upper parts, transverse ridges in the lower parts and a terminal front. A difficult task is to differentiate between rock glacier fronts and late-glacial moraines not affected or only slightly affected by permafrost creep. We addressed this issue by assessing the continuity between the terminal crest and the talus slope behind it. For moraines there is a clear discontinuity between them while for rock glaciers there is a smooth continuation to the talus slope feeding it.

Taking into consideration that rock glaciers mapping can often be highly subjective (Marcer et al., 2017), we used the methodology proposed by Schmid et al., 2015 in order to avoid overestimation or underestimation of rock glaciers occurrence. This implies mapping to be done by two different persons independently and considering only those landforms identified by both. Thus, our rock glacier inventory was intersected to that of Onaca et al., 2017a kindly provided by the author. At the end, there were considered only those landforms identified in both inventories.

3.2 GIS analysis

ArcGIS software was used to compute several morphometric parameters of each rock glaciers, i.e. surface, length, width, altitude and orientation (including of the source rockwall area). Mean values of the pixels were considered for altitude and direction

while for length and width the average value of several longitudinal and cross profiles respectively were used (Kellerer and Pirklbauer et al., 2012). Mean annual air temperature (MAAT) was also computed using a thermal lapse rate of 0.57 °C (Săvulescu, 2014, p. 96) and a temperature of –2.2 °C (average in the 1941–2017 time interval) for the altitude of 2504 at the meteorological station Vârful Omu from Bucegi Mountains, located at 40 km east from Roșu Peak. The digital elevation model that was used for all calculations was obtained from topographic maps 1:25000 with contour lines interval of 10 m.

3.3 Thermal measurements

3.3.1. Temperature of alpine springs during late summer/early autumn

The water of alpine springs has multiple sources, e.g. deep underground water, rain and snow. If permafrost occurs, ice can be an additional source that alters significantly not only the temperature of spring but also its geochemistry (Williams et al., 2006; Harrington et al., 2017; Kürty et al., 2017). Thus, several studies indicated that spring water temperature indicate possible permafrost if temperature is between 1 and 2 °C or probable permafrost if temperature is <1 °C (Frauenfelder et al., 1998). Other studies revealed that active rock glaciers have spring water temperatures below 1 °C for the entire warm season (Krainer and Mostler, 2002). Springs temperature can also be used to model potential permafrost distribution in mountain areas (Carturan et al., 2016).

We measured the temperature of 20 springs from the alpine area of Iezer Mountains in 20–21 September 2018 at the end of the warm season. Air temperature was well above 0 °C even in the days and nights before the survey. Most of the measurements were performed at the very point of the spring location but for some of those rising from scree areas the measurements were done at a certain distance from the source (meters to tens of meters). Springs are located both on valley floors and mountain slopes and have multiple outcoming sources such as meadows, regolith and debris accumulations. An Extech handheld thermometer (TM20 model) with accuracy of ± 1 °C was used for

the measurements. The instrument was calibrated in alpine springs that were previously monitored with higher precision instruments (iButtons) and offered similar results with less than 0.3 °C differences.

3.3.2. BTS

BTS is amongst the most used methods for mapping and modelling the potential permafrost distribution in cold regions around the world (Ardelean et al., 2015; Blikra and Christiansen, 2014; Colucci et al., 2016; Seppi et al., 2015). The method is based on the finding that negative thermal anomalies in the underground induced by permafrost occurrence can be detected by measuring the temperature at the snow–ground interface at the end of the winter. Probable, possible and improbable permafrost are related to sites where BTS is below –3 °C, between –2 and –3 °C and above –2 °C respectively (Haeberli et al., 1973). A prerequisite for the relevance of this method is the snow cover of at least 60 to 100 cm depth installed sufficiently early in the winter to induce thermal stability. Itinerant BTS consists in measuring the temperature using a thermal probe of a few meters long by pushing it through the snow pack down to the ground level. Some authors apply the BTS concept to the continuous temperatures measured by in situ data loggers placed in the ground (Gądek and Kędzia, 2008; Krainer et al., 2015). We refer to this method as ground surface temperature (GST) measurements and the average temperature at the end of the winter as stationary BTS in comparison to itinerant BTS described in this chapter.

BTS investigations were performed in Colților Valley on a complex of landforms below the Păpușa Peak (Site 1 – Păpușa), below the Tambura Peak (site 2 – Tambura) and in the lower part of Colților Valley (site 3 – Colților). Site 1 corresponds to a talus slope and scree located down valley with two ridges of uncertain (moraines or structural?) origin with little probability of permafrost creep origin. Site 2 is related to a rock deposit that could be a rock glacier and site 3 on a “certain” rock glacier (both described in detail in the Results section). We performed BTS measurements in March 2012 on all three sites and in March 2018 at Tambura and Colților sites.

3.3.3. GST

The method consists in measuring the ground surface temperature continuously using miniature data loggers installed in the field that measure and store the temperature at specific time intervals chosen by user. The method was initiated in the '90s in the Swiss Alps (Hoelzle et al., 1999) and is largely used in permafrost studies since then. It allows not only to infer more precisely the permafrost probability at a specific location but also to understand and quantify the complex heat exchange processes at the ground–atmosphere and ground–snow interfaces during the warm and cold season respectively (Ishikawa, 2003; Ferreira et al., 2017).

GST was monitored at Tambura site in the upper part of the blocky deposit at 2162 m asl between 2012 and 2018 (T–1) with a sampling frequency of 4 hours using miniature digital data loggers (iButtons) manufactured by Embedded Data Systems (USA). The sensor was put at the upper part of a block into a 2 cm hole, sealed with sanitary silicon and covered by a 10 cm rock fragment. A data gap between 2016 and 2017 was caused by logger failure. In 2017 – 2018 an additional logger (T–2) was hung between the boulders about 1 m deep, in the same location like T–1 and another one (T–3) was placed in the same way in the lower part of Tambura scree at 2046 m asl (Fig. 1). The declared logger accuracy is ± 0.5 °C but judging by the values registered in the snow melting period (zero curtain), the error was not larger than 0.125 °C near 0 °C (Gubler et al., 2011). The resolution of the loggers is 0.0625 °C.

The annual regime of GST was divided in five intervals in accordance to permaNET handbook recommendations (available at www.permanet-alpinespace.eu): phase 1 of ground temperature oscillations quasi identical to air temperatures; phase 2 of ground cooling caused by negative air temperatures and overcooling because of thin snow layer albedo and long wave radiation emission; phase 3 of attenuated oscillations of ground temperature imposed by a thicker snow layer; phase 4 of complete ground isolation from atmospheric temperatures imposed by a thick snow cover that will determine GST tendency to reach an equilibrium state (4A interval) usually achieved at the end of the winter (4B interval); phase 5 of snow melting when

thermal regime enters the zero curtain period. Based on data measured in 2012–2016, we calculated for each phase the average period of occurrence, duration and mean temperature.

4. Results

4.1 Rock glaciers occurrence in Iezer Mountains and their characteristics

Our inventory (I–1) comprises 17 rock glaciers while the inventory of Onaca et al., 2017a (I–2) contains double (34 landforms). 15 rock glaciers are commonly identified in both inventories with some differences for their outlines and one landform from I–1 is considered as two separate rock glaciers in I–2. One additional rock glacier is considered only in I–1 and 17 only in I–2. Thus, 16 rock glaciers identified by both surveys are considered in this study. They are grouped in two areas located in the centre to south–west of the massif in the valleys of Izvorul Iezerului, Iezerul Mare, Groapele and Boarcășului (11 rock glaciers) and in north–east in the valleys of Colților east and west, respectively (5 rock glaciers) (Fig. 1). Based on morphology and general characteristics, the rock glaciers were grouped into three types:

- Type 1: Large rock glaciers, almost completely covered by shrubs located at the lowest altitudes (mean elevation between 1900 and 2000 asl) (Fig. 2a). Two subtypes can be differentiated: a) tongue shaped rock glaciers with more expressive morphology of longitudinal and transverse ridges and furrows sourcing from glacial cirques headwall. Two rock glaciers of this subtype are found in Groapele Valley; and b) lobate rock glaciers located at the base of secondary mountain ridges with lower density of ridges and furrows (Fig. 2b). Two rock glaciers in Colților West valley, one in Colților East valley and one in Iezer valley of this subtype are found;
- Type 2: Medium to small size rock glaciers located at high altitudes (mean elevation between 2146 and 2215 m asl) with a well–defined front covered mostly by herbaceous vegetation, parallel to the crest, and separated from the talus slope by a large furrow; ridges are more or less pronounced. Three rock glaciers of this type are found in Iezer valley and another one in Iezerul Mare east valley (Fig. 2c);

- Type 3: Small rock glaciers placed at relatively medium altitudes (2000 m asl) with expressive traces of past creep processes partially covered by shrubs and grass. Two rock glaciers of this type are found in Groapele and Colților West.

The remaining four rock glaciers could not be included in any group so they will be described individually from east to west. Tambura rock glacier (Fig. 2d) located in the Colților–east valley is a complex rock deposit with its roots in a suspended glacial cirque outflowing down–valley through two rock streams with a rock outcrop in between. Longitudinal furrows and the final front are well defined (Fig. 6a). Several creeping fronts are found in the upper part of the deposit and no vegetation is present there while herbaceous and shrubs vegetation occurs in the rest of the deposit. It is not a rock glacier *stricto sensu* but rather a large rock deposit with multiple source rockwalls along its length that it is affected in a certain degree of permafrost creep. Boarcășu deposit looks like a typical tongue shaped rock glacier with a small surface (c. 1.6 ha) but a relatively large length (c. 170 m) that would indicate intense past creep processes (Fig. 2e). Apparently, two generations of fronts are visible with their concavities being c. 170 m (the lower one) and 100 m (the higher one) in length. Roșu is a small (1.2 ha) but very expressive rock glacier with striking scree wrinkles located at the highest altitude from Iezer Massif (2329 m asl). The largest furrow is more than 10 m deep and all ridges are colonized by vegetation and even shrubs. The latter are present in spite of high elevation because of south–eastern exposition. Groapele 4 is a probable rock glacier with an intermediate state between type 1 and type 3 with a medium surface and less expressive morphology completely covered with shrubs. The supplementary landform identified only in I–1 in the upper sector of Piscanu valley is considered to be a rock glacier of type 2 (Fig. 2f).

The rock glaciers from Iezer massif (hereafter referred to as the 16 features commonly identified in both I–1 and I–2) have a mean elevation varying from 1919 to 2329 m asl., mean slope between 11 and 24° and surface between 0.7 and 9.5 hectares. The average values for altitudes, slope and surface are 2056 m asl, 20° and 3.5 hectares respectively.

Taking into consideration altitude, slope and direction, the MAAT was computed for every rock glacier according to Stanciu (1981) and Iancu (1982) both cited in Săvulescu (2014, p. 49 and p.96). Values at mean elevation for each rock glacier varied between -1.2 and $+1.1$ °C with an average of $+0.4$ °C. Roșu rock glacier has a MAAT of -1.1 °C, 4 rock glaciers from Izvorul Iezerului and Iezerul Mare east have negative MAATs, Tambura and Boarcășu rock glaciers have MAATs of 0.1 °C and the other rock glaciers have MAATs of $0.6 - 1.1$ °C. All landforms with MAATs below 0.1 °C present similar geomorpho-

logic characteristics like openwork structure, vegetated ridges and terminal front while the rest of rock glaciers are covered by herbs and shrubs in a larger degree but still present a partial openwork (matrix free) structure especially on furrows and micro depressions. Rock glaciers directions are north east (6), north and north-west (3/3), east and west (1/1) and south-east (2).

The lithology of rock glaciers is represented by mica schists with garnets (13 rock glaciers), mica schists with garnets and amphibolite (two) and two mica paragneiss (one).

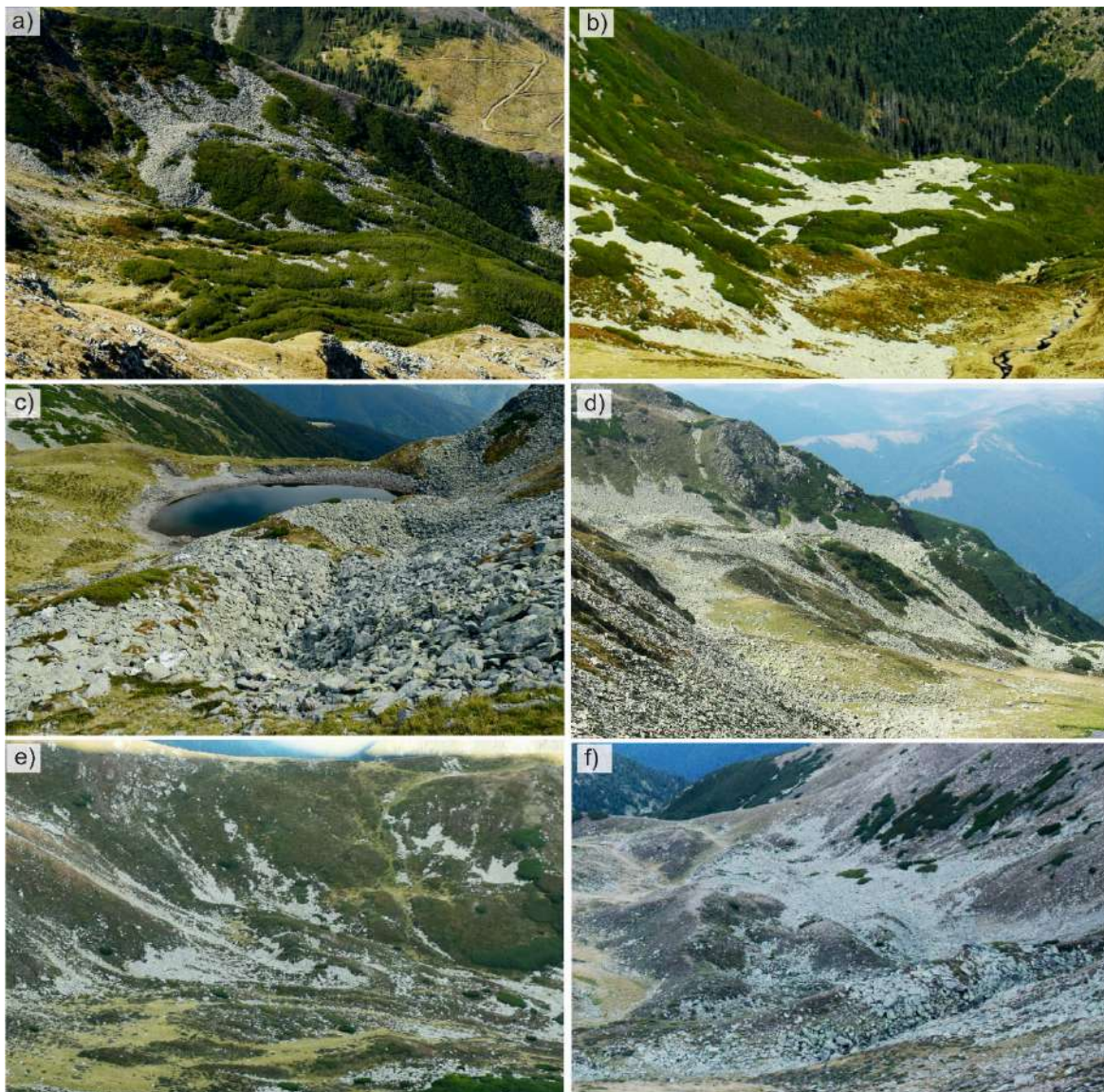


Figure 2 Rock glaciers types in Iezer Massif: Type 1a (a) and type 1b (b) rock glaciers from Groapele and Colților valleys; type 2 rock glacier from Iezer Valley close to the Iezer Lake (c); Tambura (d), Boarcășu (e) and Piscanu (f) landforms interpreted as atypical rock glaciers

4.2 Thermal characteristics of rock glaciers

4.2.1. Alpine spring temperature during late summer

The investigated springs are located at altitudes between 2050 and 2250 m asl and registered temperatures between 1.9 and 5.9 °C (Fig. 3). According to their position and source, the springs were classified into the following categories:

- Type 1 (2 cases): springs rising from slopes covered with meadows close to the mountain ridges and peaks (Păpușa South and Izvorul din Plai) with

altitudes of 2183 and 2250 m asl and temperatures of 3.4 and 4.6 °C (Fig. 4a).

- Type 2 (7 cases): springs rising below talus slopes, protalus ramparts or rock glaciers. They have altitudes between 2058 and 2172 m asl and temperatures between 1.9 and 5.3 °C. These springs have the lowest minimum and average temperatures. Two springs with temperature below the threshold of possible permafrost (2 °C) were measured in the Colților and Izvorul Iezerului valleys (Fig. 4b).

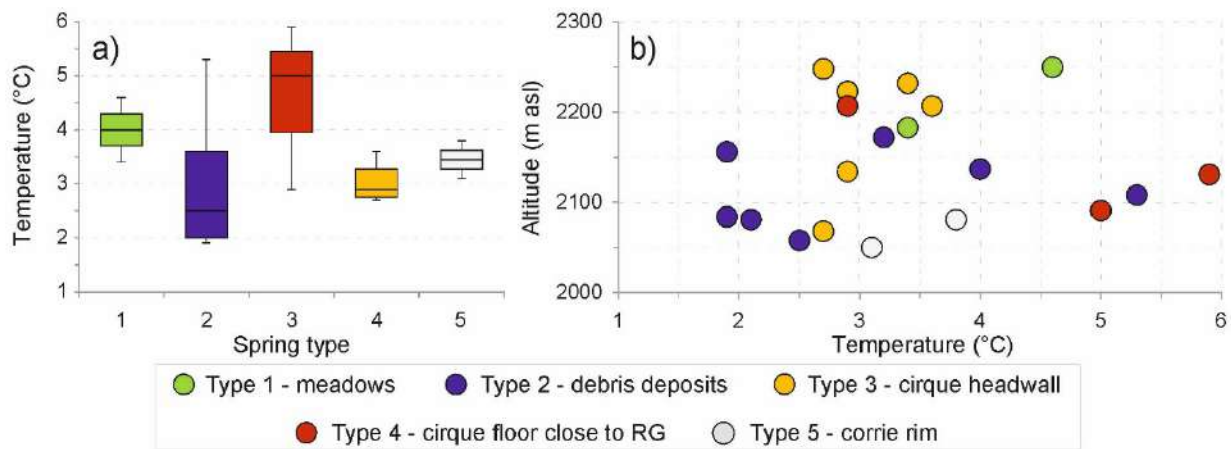


Figure 3 Temperature of alpine springs from Iezer Massif measured in 20–21 September 2018. Box plots of temperature variability of spring types from 1 to 5 (a) and temperature versus altitude (b). Spring types are discriminated by source as alpine meadows (1), screes (2), cirque floor close to rock glaciers (3), cirques headwall (4) and corrie rim (5) and are expressed by the same colour from plot a – c



Figure 4 Three examples of springs measured in the alpine area of Iezer Massif and their temperatures in relation to their source. Type 1 spring below Păpușa Peak rising from alpine meadow (a), type 2 springs in probable relation to talus slopes uphill in Colților Valley (b) and type 3 of springs rising from cirque headwall from Iezerul Mare East valley (c)

- Type 3 (6 cases): springs rising on the cirque headwall covered with more or less consistent scree deposits are located at altitudes between 2068 and 2248 m asl and have temperatures between 2.7 and 3.6 °C. Most of them are located in the lower part of the headwall and present the highest apparent dis-

charges. Such springs were measured in the Colților West, Boarcășu, Iezerul Mare East and West and Izvorul Iezerului valleys (Fig. 4c).

- Type 4 (3 cases): springs rising on the cirque floor close to rock glaciers have altitudes between 2091 and 2207 m asl and present temperatures

between 2.9 and 5.9 °C. Their lateral position to rock glaciers makes it unclear in what degree their thermal regime is related to rock glacier. Such cases are related to Tambura, Crucea Ateneului and Iezer 1 rock glacier.

- Type 5 (2 cases): springs rising at the lower limit of the glacial cirque (corrie lip) are located at altitudes of 2050 and 2081 m asl and have temperatures of 3.1 and 3.8 °C. This type of springs was measured in Boarcășu and Izvorul Iezerului valleys, the former is found below glacial and periglacial debris deposits while the latter is positioned below a rock glacier.

From a thermal point of view, relatively cold spring (two cases), below thermal threshold of permafrost presence (2 °C) can be found only below some talus slopes and rock glaciers, moderately cold springs (2 – 3 °C, seven cases) exist below some talus slopes and protalus rampart, lateral to rock glaciers and on the cirque headwall, moderately warm springs (3 – 4 °C, six cases) located at the cirque limit (corrie lip), on the cirque headwall, on the alpine meadow and below protalus rampart, and warm springs (4 – 5.9 °C, five cases) below warm

talus slopes, alpine meadow and at a lateral position from rock glaciers.

4.2.2. BTS investigations

Investigations at Păpușa site from March 2012 revealed two areas of low BTS, in the east and west corresponding with the lower sectors of talus slopes (Fig. 5). If we consider the classic BTS thresholds used for prediction, possible and probable permafrost occurs at 18 and 10 points respectively representing 44 and 24 % from the total measured points (Table 1). In 2012, measurements at Tambura site revealed possible and probable permafrost on 67 and 56 % from the total number of points (Fig. 6). In contrast, the BTS values corresponding to possible and probable permafrost were much fewer in 2018, i. e. 26 and 5 % respectively (Table 1). Colților rock glacier revealed possible and probable permafrost for 50 and 29 % of the measured points in 2012 in comparison to 11 and 0 % in 2018 (Fig. 7). At both sites, the averages and the minimum values of BTS were much higher in 2018 in comparison to 2012 the largest difference being registered at Tambura (Table 1).

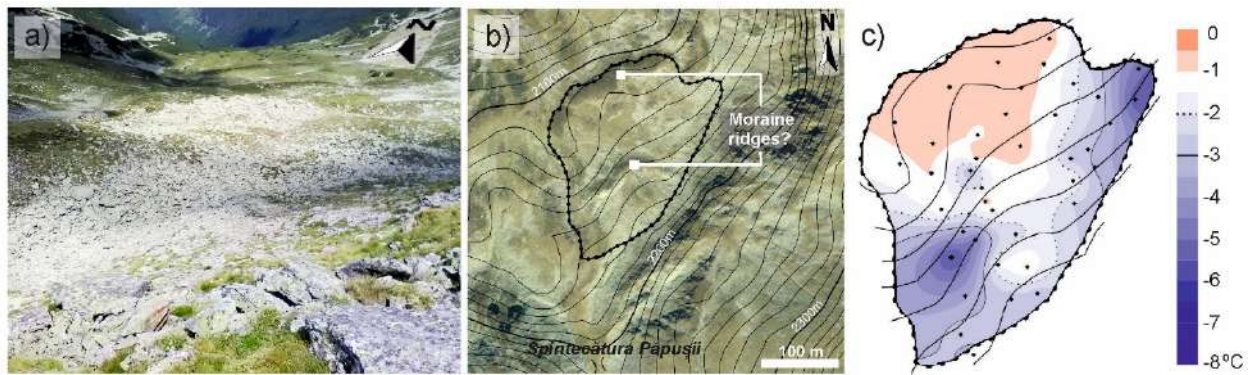


Figure 5 BTS investigations at Păpușa site from the upper part of Colților Valley from March 2012 indicating possible and probable permafrost in the upper part of the area at the contact to talus slope. The two ridges from the deposit might be rather small moraines than ridges induced by past permafrost creep

Table 1 Results from BTS field campaigns in Iezer Massif in the 3 sites from Colților Valley in terms of BTS values and predicted permafrost

Site	Area (ha)	Year	Min. BTS	Avg. BTS	Possible permafrost No. points	%	Probable permafrost No. points	%
Păpușa	4.6	2012	-5.7	-2.1	18	44	10	24
Tambura	7.2	2012	-9.3	-3.4	38	67	32	56
		2018	-3.2	-1.8	11	26	2	5
Colților	6.1	2012	-5.3	-2.1	17	50	10	29
		2018	-2.3	-0.9	4	11	0	0

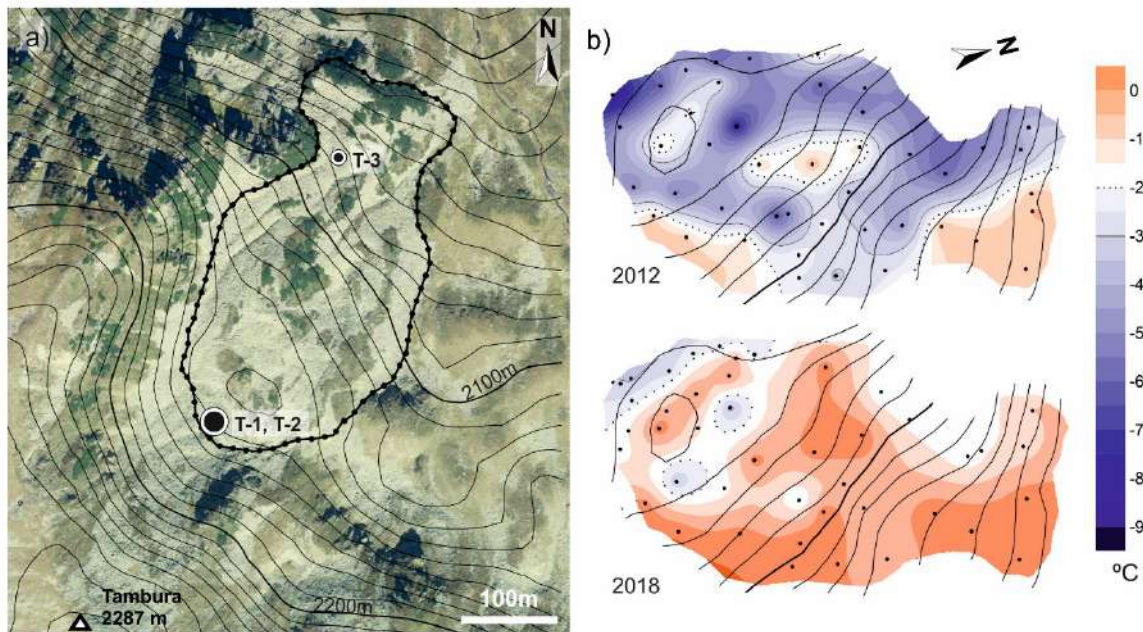


Figure 6 BTS investigations at Tambura site indicating probable (2012) and possible (2018) permafrost. The main rock outcrop is covered by shrubs and grass and the two rock streams present longitudinal furrows possibly induced by permafrost creep. Location of GST loggers is also shown in plot a

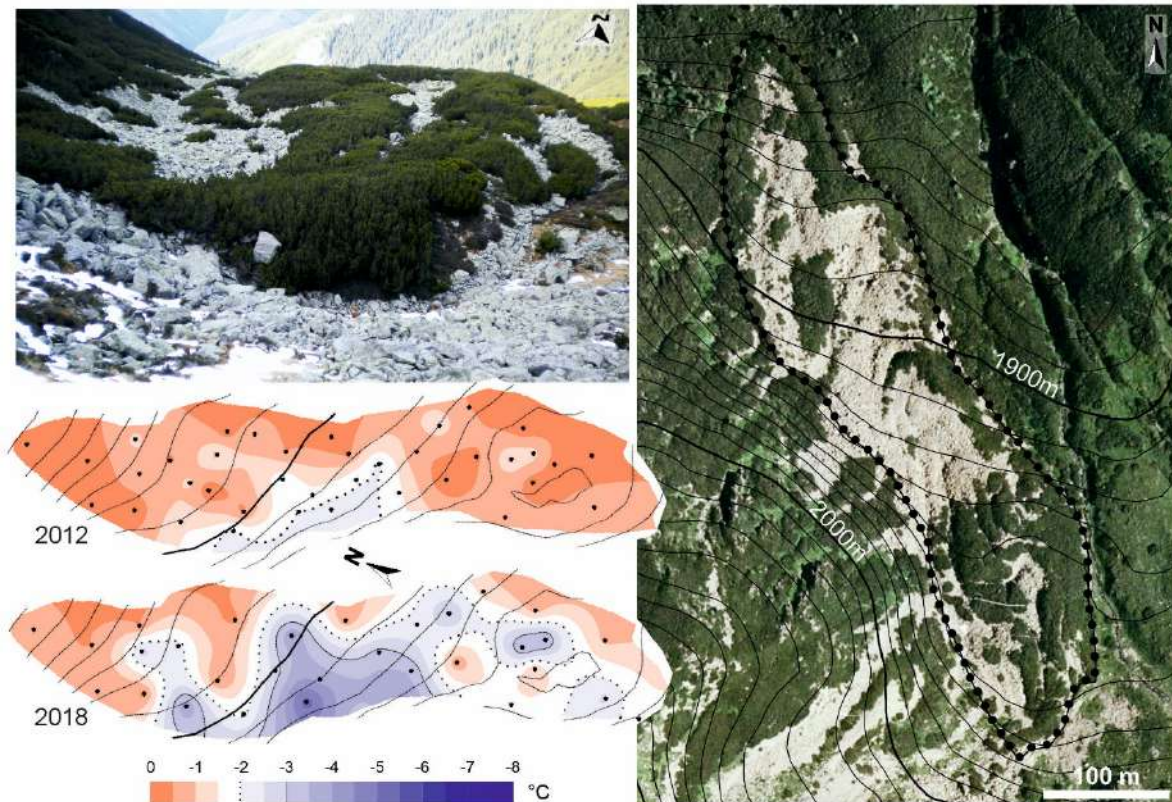


Figure 7 BTS investigations at Colților site from the middle part of Colților Valley from March 2012 and March 2018 indicating possible (2012, 2018) and probable (2012) permafrost in the middle part. Landform morphology clearly indicated past permafrost creep processes

Table 2 The average characteristics of the GST phases in terms of timing, duration, temperature and solid precipitations specific to Tambura scree possibly underlain by permafrost

GST phase	Avg. time interval		Avg. duration (days)	Avg. temp. (°C)	Avg. solid prec. (mm)
	Init.	End			
Phase 1	8-Jun	6-Nov	152	8.4	
Phase 2	6-Nov	20-Nov	17	-2.2	4.9
Phase 3	21-Nov	15-Dec	23	-4.1	48.2
Phase 4A	16-Dec	17-Feb	64	-3.4	152.2
Phase 4B	18-Feb	21-Mar	30	-2.0	75.3
Phase 5	22-Mar	10-Jun	83	–	–

Table 3 Methodology to define BTS interval. Timing and thermal oscillations are calculated for both BTS 1 and BTS 2. BTS 1 is defined as the longest interval without oscillations before the snow melting initiation while BTS 2 is defined as the longest interval before the snow melting onset with thermal amplitude less than 0.2 °C

BTS 1	Time interval		Duration (days)	Oscillations	BTS value (°C)
	Beginning	End			
2013	3-Mar	12-Mar	9	0	-2.13
2014	12-Mar	21-Mar	9	0	-2.39
2015	19-Mar	27-Mar	8	0	-2.01
2016	19-Mar	29-Mar	10	0	-1.19
Average	11-Mar	19-Mar	8.2	0	-2.02
BTS 2					
2013	20-Feb	15-Mar	23	5	-2.16
2014	26-Feb	21-Mar	24	8	-2.38
2015	3-Feb	10-Mar	36	22	-2.36
2016	25-Feb	2-Apr	38	23	-1.25
Average	18-Feb	21-Mar	30	8.8	-2.04

4.2.3. GST

The GST data and the main thermal indices are presented in Figure 8 and Table 2. Phase 1 (snow free interval) corresponds with the thawing season and it is the longest of the GST cycle of 136 – 156 days. Negative temperatures are possible but they are prevalent in phase 2 (ground cooling) that is characterized by thin and discontinuous snow cover that diminishes the amplitudes of ground thermal oscillations; the average solid precipitations are 5 mm and its duration varied between 9 and 25 days. In phase 3 (incomplete isolation) ground temperatures have no diurnal oscillations and follow air temperature only with time lag and lower amplitudes. The transition to this phase is marked by a snowfall event of >10 mm of solid precipitation, its length is highly variable (5 – 61 days) and the total solid precipitation is almost 50 mm. The progression to

phase 4A (tendency for thermal equilibrium) is marked by a heavier snowfall event (20–30 mm in the first 2–3 days) and its length varies between 31 and 90 days. The phase 4B (winter equilibrium temperature or BTS phase of the GST regime) needs a special attention in analysis and interpretation because the mean temperature during this period can be used to draw conclusion on potential permafrost occurrence at a site. Thus, in order to define objectively the limits of this period, it was calculated the frequency and magnitude of oscillations by subtracting each value from the next one and there were defined two types of 4B periods: BTS sensu stricto defined as the longest time period before the snow melt onset with no oscillations (BTS 1) and BTS in a broader sense defined as the longest period before the snow melting onset with a low number of oscillations but not surpassing an overall amplitude of 0.2 °C (BTS 2). BTS 1 lasts on average 8.2

days per winter in comparison to BTS 2 which lasts 30 days; however, the mean BTS values are virtually the same (Table 3). The time interval for one thermal oscillation of 0.0625°C (which is the resolution of the data loggers) during BTS 2 is 1.6 – 4.6 days (average of 3.4 days for one oscillation), thus, we propose to consider as BTS period the interval before the snow melting onset when temperature fluctuate less than 0.0625°C per day and the thermal amplitude is lower than 0.2°C during the entire period. The mean BTS value varied between -2.16 and -2.38°C indicating possible permafrost at Tambura site in 2012–2015 and -1.25°C in 2016 indicating improbable permafrost (Table 3). In what concerns precipitation, during this interval fall as snow on average an additional amount of 75 mm. Phase 5 (snow melting or zero curtain) is the second longest period of the GST cycle with its length only slightly varying between 77 and 87 days.

In 2017 – 2018 T-1 data revealed a similar mean BTS value of -2.38°C but a significantly warmer

thermal regime during the first part of the winter (Fig. 8b). Comparing T-1 and T-2, it can be observed a limited cooling of T-2 during autumn and the occurrence of a constant thermal difference of about $0.7 - 0.9^{\circ}\text{C}$ that maintains in the BTS period (mean BTS value at T-2 is -1.51°C). T-3 registers slightly colder ground temperatures in the first part of winter but a warmer mean BTS value (-1.22°C). Also, a much longer period of winter thermal equilibrium can be observed at T-3.

The MAGST at T-1 was mostly around $+2^{\circ}\text{C}$ in the interval of 2013–2015 and increased and remained quasi stable at around the value of 3°C (Fig. 9) in the rest of the interval. It varied between $+1.6$ and $+3.1^{\circ}\text{C}$ and the multiannual mean is $+2.4^{\circ}\text{C}$. Air temperature calculated for the same altitude (2162 m asl) had a multiannual mean of $+0.9^{\circ}\text{C}$ for the same interval and varied rather differently (Fig. 9). Mean ground temperature on almost 11 months in 2017 – 2018 were $+3.2^{\circ}\text{C}$ at T-1, $+1.4^{\circ}\text{C}$ at T-2 and $+1.1^{\circ}\text{C}$ at T3.

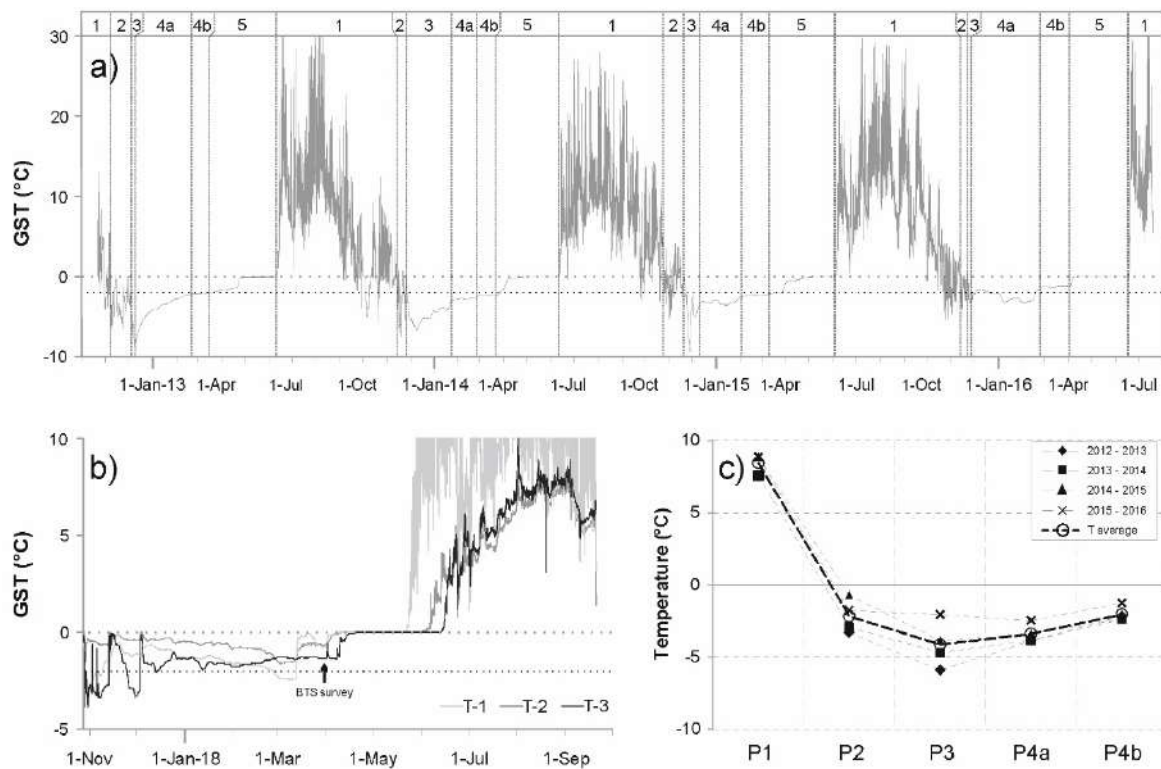


Figure 8 GST regime at Tambura site. Multiannual measurements in the upper part of the scree (T-1) at 2162 m asl with the specific annual 6 phases (a); one year GST regime at T-1 – T-3 locations with the time of itinerant BTS survey marked (b); mean temperatures of the 5 phases at multiannual level (phase 5 – zero curtain was excluded because it is close to 0°C every year) (c)

5. Discussion

5.1 Rock glaciers typology

Rock glaciers from Iezer mountains are small surface landforms (average of 3.5 ha, half of them are below this value), occurring on cirque headwalls and valley sides above the treeline. Their probable origin is mostly periglacial, i.e., talus derived. They are significantly smaller in comparison to those found in Italian Alps (5 ha) (Baroni et al., 2004), Central Alps (4.9 – 7.4 ha for talus-/moraine derived rock glaciers) (Scotti et al., 2013), Eastern Alps (6.1/7.5 ha for relict/intact rock glaciers) (Kellerer-Pirklbauer et al., 2012), Norway (10–20 ha for intact/relict and talus-/moraine derived rock glaciers) (Lilleøren and Etzelmüller, 2011), Patagonian Andes (3.6/9.2 ha for fossil/intact rock glaciers) (Falaschi et al., 2015). The main reason for such situation might be related to the type of rock glaciers that developed in Iezer Mountains which are mainly lobate talus derived rock glaciers located mostly on cirque headwalls and valley slopes. Only a few rock glaciers of type 1 occur on valley floors and have a considerable length and area (6.8–9.5 ha) in Groapele and Tambura valleys. The rest of landforms are lobate and relatively small with their surface decreasing with altitude (Fig. 10). They have a similar surface to valley wall rock glaciers (3 ha, type 2 of rock ice features) described in Sierra Nevada in comparison to cirque rock glaciers that are on average 5 times larger (Millar and Westfall, 2008). Mountain morphometry with round interfluvies and less inclined slopes did not favour the debris accumulation in large quantities but rather the spreading across large areas especially on eastern and southern slopes (subsequent and consequent). Below secondary ridges some rock deposits that formed parallel to the crest were subsequently affected by permafrost creep and formed rock glaciers of type 2. In opposition, on the western/northern obsequent slope (much steeper) rockwalls and talus slopes are considerably much better developed as well as several chaotic rock deposit of probable glacial origin. However, rock glaciers were detected just on one glacial cirque out of four in the Boarcășu basin and in two cirques out of six in Colților basin while the rest present only talus deposits under the headwall

that are partially fossilized. Such situation might be related to a reduced rockwall erosion rate in the postglacial period or a lack of debris availability in the period of climate favourability to permafrost development in Late glacial.

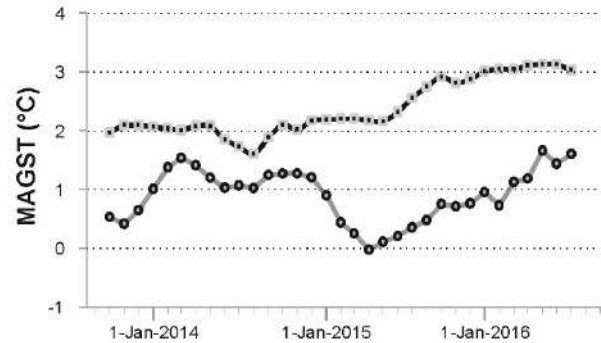


Figure 9 Multiannual variation of MAGST at Tambura site (grey squares line) along MAAT (black circles line) adjusted to the level of 2162 m computed using data from Vârful Omu meteorological station (2504 m asl) with a lapse rate of 0.57/100 m. Curves are moving averages and each point represents the mean value of the 12 previous months

The prevalence of northern direction of rock glaciers is typical because of the colder microclimatic conditions more favourable to permafrost development and creep (e.g. Kellerer-Pirklbauer et al., 2012; Scotti et al., 2013). In terms of grain size and texture most rock glaciers present medium blocks with diameters in the order of decimeters mostly matrix free that would include them in the category of bouldery type (Matsuoka et al., 2005). Grass and shrubs are embedded on convex ridges while lichens colonize the exposed parts of all blocks which indicate their lack of movement. However, this does not exclude the possibility to host permafrost, as talus rock glaciers are much more efficient in maintaining permafrost below the rock mantle in opposition to debris rock glaciers (Lilleøren and Etzelmüller, 2011).

The current inventory takes into account 16 rock glaciers, a much lower number in comparison to the previous works (Onaca et al., 2017a; Szepesi, 1998–1999). From the supplementary 17 landforms of I–2, we consider that seven are „certainly” misinterpreted as rock glaciers and are in fact scree slopes (1), lower talus slopes (2), moraine ridges (1), moraine deposits (1) and debris accumulations in cor-

respondence to structural rock outcrops (2). Two additional landforms are interpreted as incipient rock glaciers, i.e. protalus ramparts, a different landform type. For 8 landforms identified in I-2 some potential permafrost connection might exist because one morphological element that could be related to permafrost creep can be identified: an apparently well-developed front of deposits located below talus slopes (four cases), a chaotic structure of the deposit (two cases) or several longitudinal or transverse furrows (two cases). However, the resulting landforms cannot be considered rock glaciers in our view or at least not with an acceptable degree

of confidence only on orthophotos.. Some talus deposits are rather suitable for the more general term of rock ice features (RIFs) (Millar and Westfall, 2008). Such landforms are indeed more widespread in Iezer Mountains but do not fulfil the geomorphological demands to be interpreted as rock glaciers in our view. Other mass wasting forms included in RIF category like solifluction lobes are well represented in Iezer Massif and they seem to be active today. However they are only associated to seasonal frost and freeze-thaw action and not to permafrost (Fig. 11) (Onaca et al., 2017b).

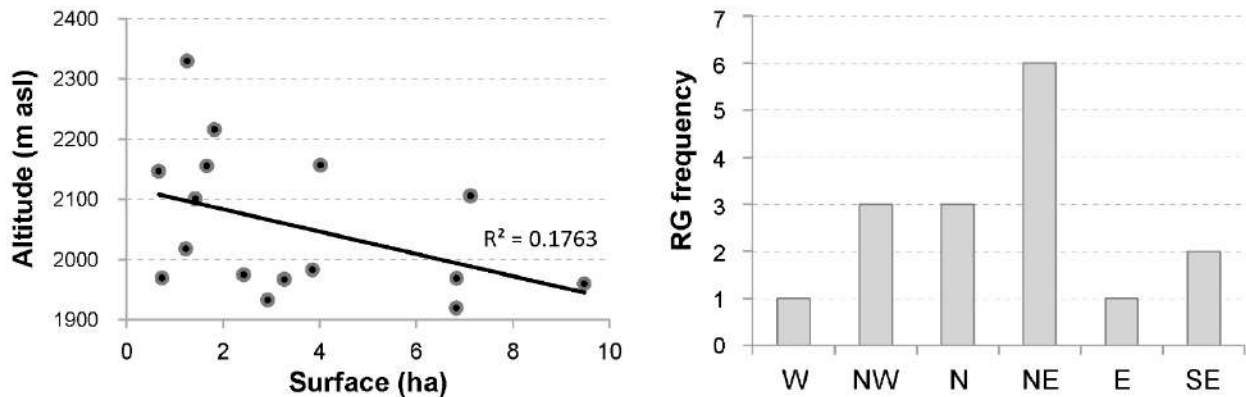


Figure 10 Rock glaciers characteristics and distribution. Relation between altitude and surface (a) and frequency on each cardinal direction (b)



Figure 11 Rock Ice features (RIFs) from Iezer Massif different from rock glaciers. Solifluction lobes on the western slope of Păpușa Peak related to seasonal frost (a) and a probable protalus lobe possibly related to incipient permafrost creep in the upper sector of Groapele Valley. The blue stick from picture a) is about 1.2 m high

Rock glaciers mapping in marginal periglacial mountain environments may be a difficult task as several problems arise from tracing the limits of the landforms especially in the areas covered by thick vegetation, differentiating them from other rock deposits like moraines, debris slopes and rock slope failures. Recent studies proved that some previously assumed rock glaciers were in fact other landforms like rock slope failures and debris drapes (Jarman et al., 2015) or “bedrock with clusters of locally derived boulders” and glacial deposits and solifluction lobes affected by fluvial processes (Wilson, 2011). This cancels previous palaeoclimatic inferences made on the basis of landforms interpreted as rock glaciers.

5.2 Thermal signature and possible permafrost occurrence

Spring water temperatures registered in the Iezer Mountains are more homogeneously distributed in comparison to Retezat (Vespremeanu–Stroe et al., 2012) and Parâng Massifs (Urdea and Vuia, 2000; Popescu et al., 2015). Also, their minimum temperatures are slightly higher. We assume that springs located on cirque headwalls, corrie rim and bellow some talus deposits have the typical temperatures for such a mountain environment, between 2.5 and 4.5 °C. Those with more vigorous discharges might indicate a better connection to the underground faults system and have probably a more stable thermal regime as well. Yet, three atypically cold springs were found in Colților and Izvorul Iezerului Valleys where temperature was 1.9 ... 2.1 °C, i.e. around the threshold of possible permafrost. They are related to a talus slope located 200 m uphill below the Păpușa Peak, and below a rock glacier front in the close vicinity of Iezer Lake. These two structures have a certain probability to host some low volume late summer frozen structures. Springs located in the lateral parts of rock glaciers are probably not connected to their runoff system and that is why their temperatures are so high. The warmest springs (above 5 °C) are among the weakest in terms of discharge and their temperatures could be related to warming caused by low flow velocity and enhanced heat transfer. That suggests that applying this method so late in the autumn could alter the results because of too low discharges. Our results confirm the investigations performed by Szepesi,

1998–1999 who suggests alpine springs temperatures below 2 °C in the summer.

Spring temperatures from Iezer Mountains are significantly warmer in comparison to other massifs where ice containing talus and rock glaciers lies in the vicinity of glaciers (Carturan et al., 2016). Also, the water from Iezer springs is clearly indicating motionless talus with no permafrost creep involved in comparison to other sites where active rock glaciers and rock ice forms present streams with suspended silt (Millar and Westfall, 2008). Also, active rock glaciers have lower temperatures (below 1.3 °C) for the entire warm season and their discharges are highly variable (Krainer et al., 2015). Because of missing massive ground ice and the presence of sandy to silty basal layer, relict rock glaciers were proved to have a great storage capacity acting as long term (in the order of months) aquifers with a regulating effect on hazards like floods and debris flows from mountain areas (Winkler et al., 2016). That is even more important as they are landforms with a long residence time in the landscape in comparison to glacial landforms (Knight et al., 2019).

BTS investigations at Păpușa site indicate two overcooled scree areas that might be related to the two cold springs from Colților Valley (Fig. 4b). This situation of low temperature springs rising tens to hundreds of meters from a potential frozen source is similar to that from Doamnei rock glaciers (Făgăraș Massif). Potential permafrost areas at Tambura and Colților sites vary significantly from 2012 to 2018 (Table 1). The GST regime proves however that this second BTS survey (31 March 2018) was performed at a time when BTS period was finished at T1 and T2 sensors locations and the temperature was 0.5 – 1.5 °C higher than in the BTS interval (Fig. 8a). Still, at T3 location, BTS period finished 5 days later. These facts indicate that BTS maps from 2018 are not completely relevant. An average situation, at the middle of the two estimations might be more appropriate because 2011–2012 winter induced atypical ground conditions as well but in the sense of a more intense cooling (Popescu et al., 2015).

GST measured in the upper part of Tambura scree (T–1) confirmed the potential permafrost occurrence suggested by BTS, with the exception of 2015–2016 when BTS was higher than –2 °C. The

latter was caused by the short durations of phase 2 and 3 and the early initiation of phase 4a before the ground cooled. In fact, the warmer temperatures in phase 2 and 3 are in power to increase the mean BTS value even though the opposite does not occur at our study site (Fig. 8c). A moderate correlation between ground freezing index and mean BTS was found for rock glaciers in Retezat Massif (Onaca et al., 2015). It was also shown based on borehole data that the more snow in the early winter the higher the temperature in March will be at the top of mountain permafrost (Vonder Mühll et al., 1998).

In 2017–2018 interval, GST at 1 m depth (T2) was warmer than at the surface (T1) for the entire winter especially at the beginning of monitoring period indicating a reduced ground cooling efficiency in spite of blocky layer. The thermal difference of 0.7 – 0.9 °C during most of the winter indicates a lack of air stratification below the insulating snow cover. However, the warmer temperatures at T2 in the BTS period do not indicate permafrost absence. Temperature increase with depth was also visible at other rock glaciers underlain by permafrost. For example in Swiss Alps Murtél rock glacier, temperature measured in boreholes in 1 April 2000 increased from approximately –4 °C at the surface to –2.7 °C at –1.5 m and –1.7 °C at –5 meters or in the Muragl rock glacier from –2.8 °C at the surface to –2 °C at –1 m and –1 °C at –2.5 m (Vonder Mühll et al., 2003). Thermal investigations in shallow boreholes in Scandinavia revealed the temperature increase with depths both in permafrost and seasonal frost areas and during the entire winter at sites with continuous winter cold air infiltration in the ground through a thin snow cover (Juliussen and Humlum, 2008).

MAGST values similar to those recorded at T1 in correspondence with BTS >-2 °C were attributed to relict rock glaciers in Retezat, Parâng and Făgăraș Massifs (Vespremeanu–Stroe et al., 2012; Onaca et al., 2013; Popescu et al., 2015) but in other marginal periglacial massif (Atlas, North Africa) it was suggested that they can still be attributed to permafrost if associated with BTS below –2 °C (Vieira et al., 2017). Positive MAGST (0.8 – 2.8 °C) was also found in a relict rock glacier subjected to probable permafrost in its rooting zones in Austrian Alps at a mean

annual air temperature of >4 °C (Kellerer–Pirklbauer et al., 2015) and creeping permafrost at positive MAGST (1–2 °C) in Italian Alps. In the latter case, however, rock glacier is derived from a former glacier (Seppi et al., 2015). The average surface offset (MAGST – MAAT) in 2013–2016 was 1.5 °C which is identical to that of Murtél rock glacier (Hoelzle and Gruber, 2008).

Taking into consideration that possible permafrost was documented at 1900 m asl in Colților rock glacier and that openwork coarse blocks can assure the thermal conditions for permafrost maintenance even at unexpectedly low altitudes, we suggest that most of rock glaciers (11) from Iezer Massif considered in this study are underlain by patches of permafrost and could be framed in the class of inactive rock glaciers. This assumption is supported by the findings of Kellerer Pirklbauer et al. (2005) who documented probable permafrost in a rock glacier located in conditions of MAAT of 1.6 °C in the eastern Alps. The causes of such low altitude distribution might be related to the openwork structure and porous debris availability and to snow cover conditions which is also crucial. Thermal processes in the blocky layer like funnelling through thin snow cover and hoarfrost crystal formation decrease significantly the MAGST (Bernhard et al., 1998; Kellerer Pirklbauer et al., 2015). However, we suggest that inactive rock glaciers from Iezer Massif are much closer to relict rock glaciers in terms of isolated permafrost occurrence in small spots and subdued morphology in comparison to those considered inactive in the Alps which seem to be much closer to active rock glaciers, presenting permafrost on their almost entire area but in a motionless state. Relict rock glaciers are considered those from Groapele Valley which are fully covered in shrubs and the one from Boarcășu cirque which has a matrix supported debris deposit of probable moraine origin (Onaca et al., 2017a). Moreover, we suggest that sporadic patches of permafrost might exist in other debris deposits and rock ice features like talus and scree slopes of non-glacial origin and with openwork structure. This is significantly more than previous assumptions of Onaca et al., 2017 who indicated the presence of 4 intact rock glaciers in Iezer Massif or from Szepesi, 1998–1999 who suggested that only

Roşu rock glacier might be inactive based on its altitude. In fact, this rock glacier might hold the record for the highest from Romanian Carpathians with an average altitude of 2329 m asl.

Geomorphological aspect of debris deposits in Iezer Massifs including rock glaciers do not show a recent acceleration of dynamics as it was documented in the Alps, they seem well settled while other mass movement processes like debris flow can be noticed to increase in occurrence. That is because coarse and porous deposits can assure great thermal anomalies at depths that protects efficiently the underground as it was documented at Murtél rock glacier for example (Schneider et al., 2012).

6. Conclusions

A lower number of rock glaciers (16) is assumed to occur in Iezer Massif in comparison to previous estimations based on two inventories independently produced. In mapping, we considered only landforms fulfilling at least two morphometric parameters to consider a rock glacier, i.e. ridges and furrows topography and a visible terminal front. Other eight additional landforms might be rock glaciers but field visits need to be done to check for the both criteria. Several other landforms that could be included in the more general term of rock ice features some affected by permafrost creep, can be found in Iezer Massif. Such landforms like protalus ramparts, protalus lobes, boulder streams and solifluction lobes need a better assessment, inventory and classification in the future. The small rock glaciers surface (3.5 ha on average) is related to the reduced debris production and accumulation imposed by morphometry of the valleys network but also on the relatively short period of climatic favourability.

Extensive investigations on alpine springs indicated two possible permafrost areas in a talus slope and rock glacier from Colţilor and Izvorul Iezerului glacial cirques with temperatures around the threshold of 2 °C during late September. Most of the springs are between 2.5 – 4.5 °C and only exceptionally above and below these values. We assume possible warming effects for all spring from the alpine area caused by low autumn discharges and associated lower water flow speed. Most rock

glaciers don't present a spring at their fronts, so this method cannot be used to check for permafrost presence.

BTS investigations indicate possible permafrost occurrence on Colţilor Valley down to an altitude of 1900 m which is equivalent to a MAAT of 1.1 °C. Multiannual GST proved the stability of BTS indicating possible permafrost in 5 out of 6 years of measurements. The specifics of intra-annual GST phases confirm the major role of ground cooling during autumn and early winter for inducing potential permafrost distribution in scree deposits. MAGST well above 0 °C at Tambura site indicate a large thermal gradient in the active layer above the potential permafrost. 11 rock glaciers from Iezer Massif are probably underlain by permafrost patches and could be considered inactive. This is much more than previously suggested. Other rock deposits and rock ice features like protalus ramparts and lobes, talus slopes and boulder streams located in shadowed conditions might also contain isolated patches of permafrost or at least prolonged summer ice remnants because of the cooling effect of coarse blocks and the favourable conditions of snow.

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Modelling flood hazard at Paratrechos/Peritsi stream, Naxos Island (Cyclades, Greece)

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ABSTRACT

Flash floods are one of the most significant natural hazards in the Mediterranean, affecting most basins in the region. Flash floods usually occur in ephemeral streams after particularly heavy rainfall. Although the average annual rainfall in the Cycladic islands (Greece) does not exceed 400 mm, flash floods are frequent, since most of that rainfall is often concentrated in few but very intense events. The case study is the drainage basin of Paratrechos or Peritsi stream in Naxos Island. This particular basin was selected due to the recorded flood events, the stream's proximity to the city that intensifies the flood impact, the slopes' gradient abrupt changes and the diverse infiltration rate of the geological formations. Furthermore, some parts of the basin are intact by human activities, while others are severely affected. This study investigates the relationship between rain and runoff through Unitary Hydrographs, in order to estimate and evaluate the flood potential of the aforementioned hydrologic basin and to determine the causes of the hazard and vulnerability. The Clark method was implemented and its results were further evaluated by field measurements.

KEYWORDS

flash flood; flood risk; Clark method; hazard and risk mapping; river; Greece

1. Introduction

Among natural disasters, floods are the most frequent and destructive worldwide. Floods put human lives and property at risk, destroy infrastructures as

well as agricultural and livestock facilities, induce intense soil erosion and pollution of water masses. According to Downton and Pielke (2001) and Golian et al. (2010) floods affect the greatest number of

population and provoke the most damage of all natural disasters annually, while according to Pilon (2004) floods cause one third of the estimated losses due to natural catastrophes worldwide. According to insurance organizations worldwide, floods show significant rates amongst other types of disasters both in absolute event numbers, as well as with regard to financial losses. In particular, floods represent 26% of the number of large-scale catastrophic events between 1950 and 2002, as well as 8% of deaths and 27% of the economic damage caused by natural disasters, for the same period (Munich, 2002).

In the Greek region, floods are among the most significant natural, both in financial terms and in terms of human lives losses. According to Nikolaidou and Chatzichristou (1995) 216 people lost their lives between 1887 and 1994 in Greece. Apart from life loss, floods also have a significant impact on properties (houses, shops, industries), agriculture, livestock, infrastructure (structures, roads) and utility lines. In addition, there are important consequences on the environment although the magnitude of the impact has not been adequately investigated. One of the technical peculiarities of the Greek region is the lack of instrumental data especially with regard to the systematic recording of water courses' runoff. According to Stathis (2004) in the period 1980–2002, 114 of 150 floods occurred in ephemeral torrents traversing residential areas, while the remaining events occurred in more extensive hydrographic networks. The most recent catastrophic flood event occurred on November 2017 in Mandra (Western Attica, Greece), causing many fatalities as well as material damages (e.g. Andreidakis et al., 2018).

The management of flood risk, in recent years, has focused on approaches related to flood forecasting, early warning, land use planning and production of hazard maps (e.g. Lohani et al., 2014; Alfieri et al., 2014; Cools et al., 2016). At the same time structural and mechanical measures, such as dams, canals etc., have been viewed with scepticism due to the economic and environmental disadvantages (Golian et al., 2010).

During the last decades, flooding near the limits of residential areas has been increasing; according

to Lekkas (2000) and Stathis (2004), the primary causes of this is the limitation of the streambeds (e.g. natural beds are replaced with artificial channels), the blocking of the riverbed, the decline of woodlands due to fires and the reduction of infiltration within the urban environment.

In this context, the objective of this paper is to study the phenomenon of flooding on the island of Naxos (Cyclades, Greece), assess the flood hazard and suggest actions to address flood risk. The availability of measurements concerning flood runoff is relatively limited in the Greek region and in southern Europe (Gaume et al., 2009). For this reason, hydraulic and hydrological methods have been employed for the calculation of hydrological parameters.

2. Study area

Naxos Island lies in the center of the Cycladic Islands complex with an area of 428 km² (Fig. 1). The island constitutes part of the Atticocycladic unit, and it may be described as an elliptic dome, of NE direction, which consists mainly of metamorphic schist rocks, gneiss, marbles, Neogene and Quaternary deposits. In the core of the dome lies the migmatite intrusion.

The study area is the drainage basin of Paratrechos or Peritsi stream, which is located in the western part of Naxos Island (Figs. 1, 2). The catchment area of the stream is approximately 54 km². The average elevation of the area reaches 245 m, while the average slope of the basin is 21%. The catchment area is drained by a medium to dense drainage system, which at some locations is characterized by meandering. Paratrechos or Peritsi stream is known to have caused severe problems following intense rainfall, with landslides in the mountainous areas, subsidence in large parts of the road network, and intense flooding events in the wider area of Chora, the island's capital city. This stream is very important as it is connected with the salt pit of Naxos, a well-known aquatic ecosystem on the island. Several floods have been recorded on Peritsi stream, with some recent events at the end of March 1998, beginning of January 1999, February 18, 2003, and February 4, 2011.

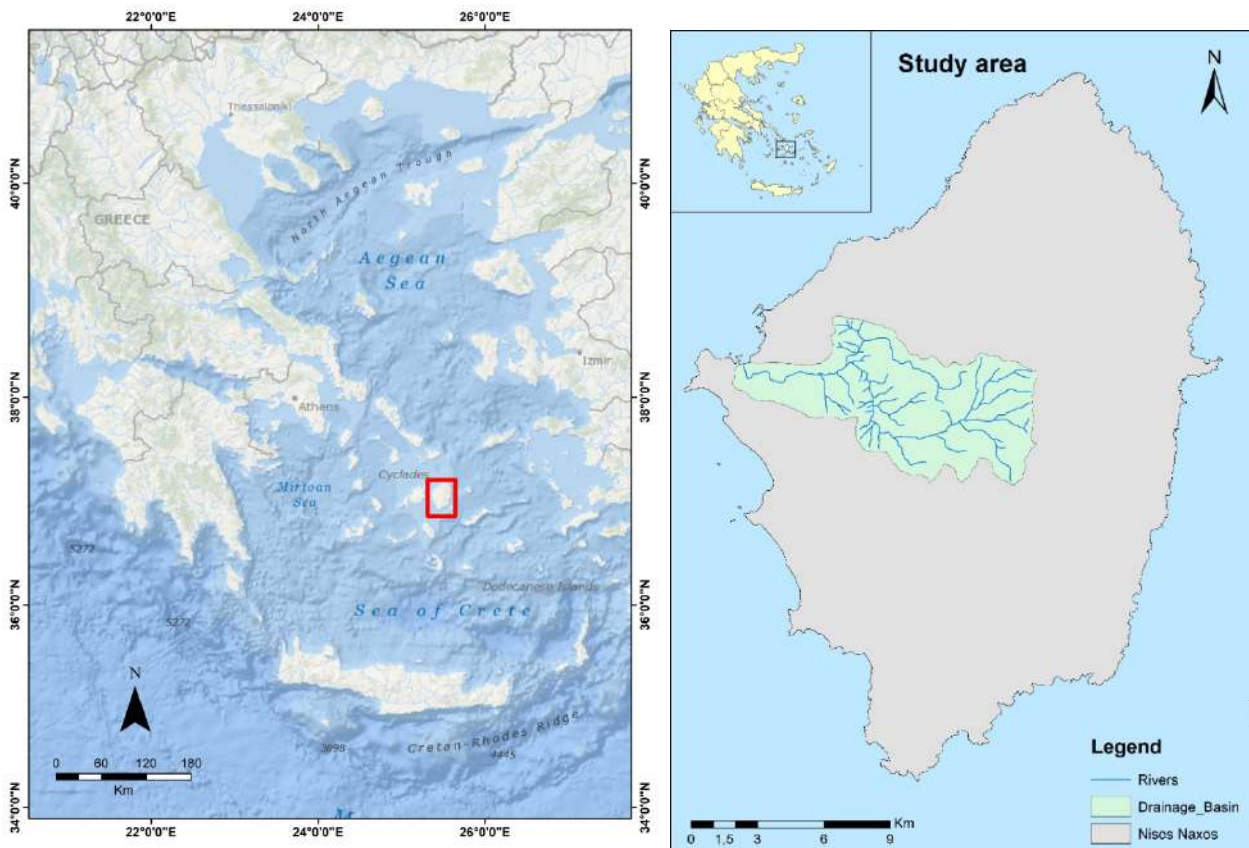


Figure 1 a) Location of Naxos Island, b) location of the studied hydrologic basin

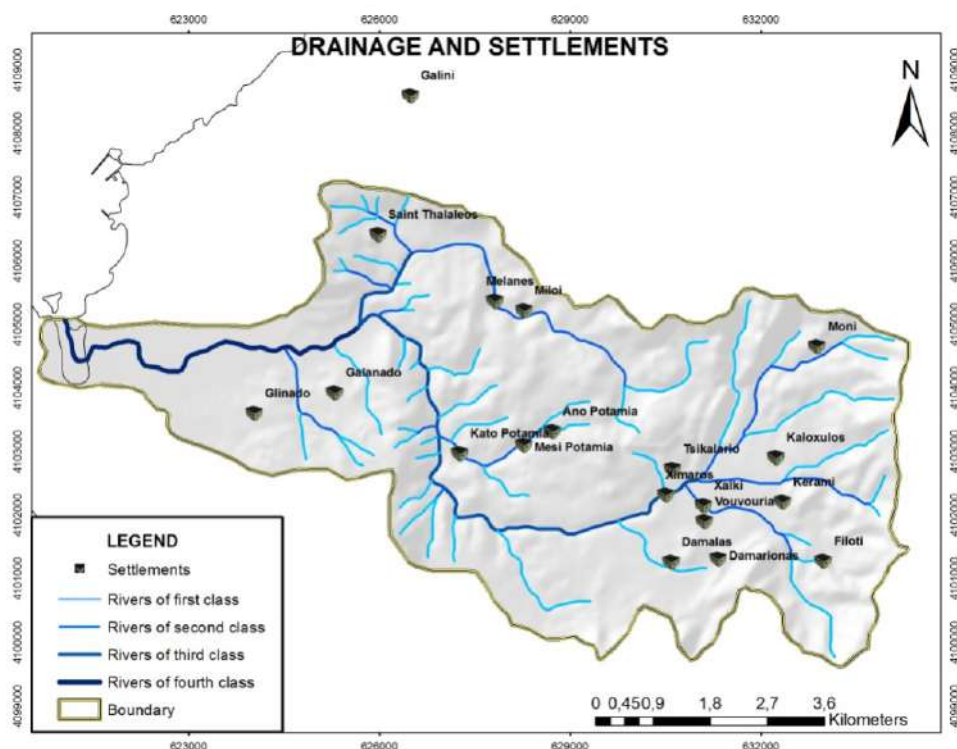


Figure 2 The studied drainage basin and the main settlements

3. Materials and Methods

For the purposes of this work, the Clark method (e.g. Clark, 1945; Yang et al., 1999; Noorbakhsh et al., 2005; Bahram, 2005; Paudel et al., 2009) was selected amongst the available hydrological models, due to the quantity and quality of the data available for the study area. As already noted, there are no satisfactory data concerning the parameters required in most other hydrological models, such as

river flows, spring discharge or groundwater levels and especially not for a sufficient timeline.

The Clark method produces Instantaneous Unitary Hydrographs (I.U.H.) for each drainage basin simulating two critical processes of transformation of excess rainfall to runoff. The maximum flow at the mouth of each sub-basin and the concentration time (the time rain water takes to reach the mouth of the basin) are calculated, in order to obtain the maximum discharge and produce the I.U.H. (Fig. 3).

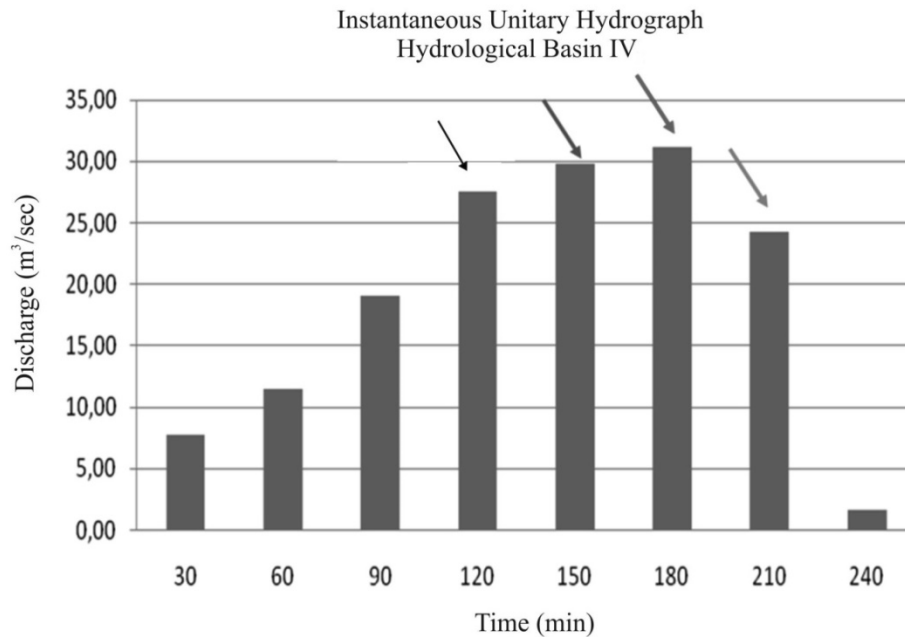


Figure 3 Instantaneous Unitary Hydrograph for drainage basin IV₁

The calculation of the flood peak was based on the Natural Resources Conservation Service parametric model unit hydrograph (NRCS). The runoff curve number (*CN*) is a coefficient that reduces the total precipitation to runoff potential, after subtracting evaporation, absorption, transpiration and surface storage for each sub-basin. The average value of *CN* for the entire basin is estimated according to the following formula:

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

where *A* is the area of the sub basin in km².

The concentration time *T_c* is the average time required for water to reach the farthest point of the basin to its orifice. In order to develop the I.U.H., the travel time grid file was divided in zones of equal time intervals, determined by the total concentra-

tion time for the whole basin. The surface of each zone was estimated and the volume was calculated, by multiplying each surface by 0.001 m, which is considered the amount of rain that falls in the catchment area, in a uniform way. The volume was used to calculate the discharge, which led to the creation of the I.U.H. (discharge in m³/sec by time in min).

The evaluation of the method was based on data derived from the installation of a hydrometric gauge and a weather station in the study area, in order to compare the peak periods of rainfall and water level. Data from three more weather stations, located within the drainage basin, were gathered and statistically processed in order to identify the period of highest precipitation and most frequently occurring flood events.

The hydrometric gauge was placed below the new Peritsi Bridge, in the downstream area because all the 4th class stream waters accumulate at this point, the riverbed is well shaped and flood events have been recorded. Measurements of the water level were collected with a recording rate of 10 min, over a period of four months (12/01/2011 to 19/05/2011), when the weather data indicated flood hazard is higher, in order to calculate the discharge. The flow velocity value that was used (1.18 m/sec) was proposed for the specific location by Stamos and Danezis (2000).

4. Results

Figure 3 depicts the I.U.H. that was developed for the stream basin (4th class–IV) of Peritsi, while table 1 shows the CN, the average slope, the length of the main branch and the concentration time, as calculated for each basin (4th class–IV and 3rd class III1 and III2).

Based on the implementation in the study area, it was calculated that the time required for water to travel from upstream to the location of the hydrometric gauge is approximately three hours (163 min), while the maximum discharge peak at that location is 40.71 m³/sec.

Furthermore, according to fieldwork measurements and rainfall data, at the bridge where the

hydrometric gauge is located, the water level rises three hours after a rainfall event upstream and the maximum discharge peak at the same location is 43.3 m³/sec (Table 2).

Hydrographs were constructed to show how water level increases at specific time periods upstream of the new Peritsi bridge, where the hydrometric gauge was placed. The 10 min time step recording of the water-level measurements was compared to precipitation events, provided by the local weather stations. These graphs are used to compute the actual time water takes to reach the point of the bridge from upstream in order to evaluate the results of the Clark method. Amongst the three flood events that were isolated (see Figure 3), the peak rainfall upstream was recorded about three hours earlier than the peak observed in the water-level measurements in the downstream region, i.e. at Peritsi bridge.

The conversion of the water level to discharge allows to compare the peak discharge from the field measurements with the one produced by the model. The greatest discharge was calculated at 43.3 m³/sec from a water level measurement recorded on February 4, 2011.

The study resulted in the production of a flood hazard map (Fig. 4), which depicts the three areas where risk is higher.

Table 1 Results of the Clark Method for the studied hydrologic basins

<i>Class</i>	Curve Number CN	Average slope (%)	Average Slope (feet/feet)	Branch length (m)	Branch Length (feet)	Tc (min)	Tc (hours)	Maximum discharge (m³/sec)
<i>IV</i>	77	21	0.21	19715	64682	213.41	3.56	31.13
<i>III2</i>	76	23	0.23	13492	44265	155.04	2.58	28.73
<i>III1</i>	80	23	0.23	9083	29800	100.14	1.67	11.89

Table 2 Measured values from the hydrometric gauge installed at Peritsi stream

Daily peaks per month		
Date (dd/mm/yyyy)	Level (cm)	Discharge (m³/sec)
28/1/2011	129	18.5
4/2/2011	241	43.3
8/3/2011	115	15.42
3/4/2011	91	10.11

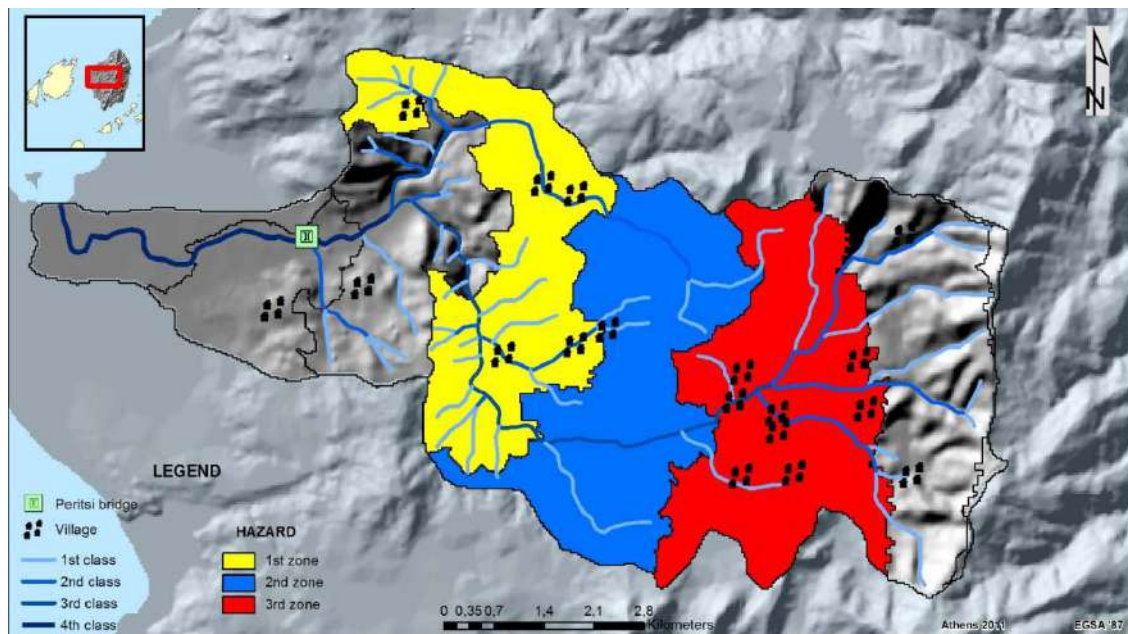


Figure 4 Flood hazard map of the study area

5. Discussion

Three main zones were distinguished (Fig. 4), whose characteristics constitute the cause of hazard at the mouth of the hydrologic basin. On the flood hazard map of Fig. 4, the three areas where risk is higher are indicated with three different color zones. A zone of very high risk (Fig. 4, 3rd zone) is identified at the mountainous area in the eastern part of the basin, while the second and third zones are areas in the central part of the basin (Fig. 4, 1st and 2nd zone). In these areas, the main elements responsible for the flood hazard at the mouth of the river basin (and for the flood peak) are the high topographic gradients, the shape of the basin, the shape of the morphological gradient, the geographical distribution of the hydrographical network and the soil (Fig. 4).

After the flood on February 18, 2003 landslides and subsidence of roads within the very high risk area were recorded, while areas located in the high risk zone caused overflow of the stream downstream. Areas downstream of the high risk area have suffered the greatest catastrophe, mainly in the first downstream area, where Peritsi bridge is located.

A comparison between the graphs of water level and rainfall has shown that, at the bridge, water

level increases three hours after the rainfall incident upstream (Fig. 2).

By comparing the results of the model and the in situ measurements it was noted that the peak discharge of the model at the mouth of the 4th class basin is identical with the one obtained from the sum of peak discharge at the mouths of the two 3rd class sub-basins.

Flood risk management in the study area requires specific and clearly defined actions. For the slopes these actions could include vegetation covers, reduction of the slope's inclination or/and height, avoiding external loads at the body and the top of the slope, retaining measures, (shotcrete, construction of retaining walls at the base of the slope, etc.). For the arrangement of the stream's bed proposed solutions may include cleaning and restoring of the existing ditches from dense vegetation, pumping of water and removal of small dams-obstacles. Finally, for the areas where flood risk is actually high a warning system is advisable to be installed.

6. Conclusions

Our study has shown that there are three main zones of flood hazard, at the study area of Paratrechos stream. The simulation of the study area and

the respective results, have led to the conclusion that the time required for water to travel from upstream to the location of the hydrometric gauge is about three hours (163 min), the maximum discharge peak at that location is 40.71 m³/sec. According to fieldwork results and rainfall data, at the bridge, the water level rises three hours after a rainfall event upstream and the maximum discharge peak at the location of the hydrometric gauge is 43.3 m³/sec. It should be noted that the results of the model were in very good agreement with in situ measurements; therefore it may be used to assess the discharge level. Last but not least, the study area has experienced catastrophic floods in the past and therefore specific and clearly defined actions are required for flood risk management. Our results have shown that the applied methodology may be used to assess the discharge level of a stream, and the acquired data may be further used to identify areas that will be affected and therefore propose and design specific actions and measures.

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Landslide susceptibility assessment using the maximum entropy model in a sector of the Cluj–Napoca Municipality, Romania

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ABSTRACT

The administrative territory of Cluj–Napoca, due to its specific geological and geomorphological characteristics and anthropic activities, has been affected for a long time by landslides. Thus, it becomes necessary to analyse affected areas with different spatial methods, with the aim of generating landslide susceptibility maps. In this research, we studied the most prone area of the city, the Becaş stream watershed, situated in the Southern part of the municipality. The aim of this paper is to generate a valid susceptibility map, to be able to raise awareness about the existing situation: due to human induced activities and rapid urban growth, the peripheral part of Cluj–Napoca becomes more and more prone to mass–movements. We used the maximum entropy (MaxEnt) model, which was fed with accurate information on the existing landslides and seven landslide–causing factors: slope, aspect, land–use, depth of fragmentation, geology and plan– and profile curvature. The results confirm that the most influential factors are the land use and slope–angle, affected in a large degree by human activities. The accuracy of the generated map was verified using the AUC method, proving a very good performance (0.844) of the applied model.

KEYWORDS

Landslide susceptibility analysis; spatial analysis; Maximum Entropy (MaxEnt); Cluj–Napoca Municipality; GIS

1. Introduction

Mass movements are a common danger; they describe the movement of rock, debris, soil, or earth material by gravity (Davis and Blesius, 2015). According to Varnes (1978), a multitude of mass movement types can be distinguished. He defines landslides as a special type of mass movement; their size can vary (smaller ones are more common) and are triggered when a threshold of stability is exceeded (earthquakes, excess of water, hydrologic inputs; Schumm, 1979; Uchida et al., 2001; Montgomery, 1994).

Landslides are amongst the most significant and dangerous geological hazards, inducing economic damage and loss, affecting the infrastructure and cities (Park, 2015; Kornejady et al., 2017).

In Romania, landslides are the most common hazards; they manifest in vulnerable areas with important human activities, like road constructions, deforestation, mining, land use change, inducing many negative effects, favoured by the absence of hazard mitigation or land planning strategies. (Kerekés et al., 2016; Necula and Niculiță, 2017). Landslides have been analysed for many years in a classical manner by Romanian researchers: Tufescu (1966), Băcăuanu (1980), Surdeanu (1988), Manea and Surdeanu (2012).

In the past decades, the interest in landslide susceptibility mapping has increased, caused by two reasons: the growing pressure of urban development and the increasing awareness of the socio-economic significance of landslides (Aleotti and Chowdhury, 1999). In order to prevent future mass movements and to generate relevant land use strategies, it becomes necessary to analyse the affected territory with different methods. After analysing the area and identifying the landslide causing factors (both natural and anthropic), a landslide susceptibility map can be generated using spatial statistics, to provide information on the expected trends in landslide occurrence in the affected area.

In the recent decades, many studies were carried out to generate relevant landslide susceptibility maps. These studies can be divided in three main categories: heuristic, statistical, probabilistic and deterministic models (Pradhan et al., 2010; Kavzoglu

et al., 2015). Several studies exist which compare and discuss the advantages and drawbacks of different methods, proving that the decision on the most relevant model to generate susceptibility maps is not a straightforward one (e.g. Guzzetti et al., 1999; Yilmaz, 2009; Felicísimo et al., 2013; Chen et al., 2016). Moreover, the selected method is suitable for cases which have specific environmental characteristics; therefore the same methodology can't be applied with the same reliability on other regions (Van Den Eeckhaut et al., 2009). In order to obtain more accurate information, it is necessary to combine two or more methods (Rossi et al., 2009).

According to Davis and Blesius (2015), heuristic methods are considered as being the basic models, especially appropriate for coarse scales; they can be easily implemented in GIS environment and are used for multi-criteria analysis or physical modelling, e.g. Blesius and Weirich, 2010, Kavzoglu et al., 2014.

Many statistic-probabilistic models were applied for landslide susceptibility, such as logistic regression, frequency ratio models, weight of evidence, random forest and maximum entropy, e.g. Luzi and Pergalani, 1999; Dahal et al., 2008; Pradhan and Youssef, 2009; Meinhardt et al., 2015; Pandey, 2015; Trigila et al., 2015; Ding et al., 2016.

Deterministic models, artificial neural network or fuzzy logic were adopted in the recent years with a great success (Neaupane and Achet, 2002; Yilmaz, 2009; Nefeslioglu et al., 2008; Kumar and Anbalagan, 2015a).

Few studies applied the maximum entropy model. This model is a statistical-probabilistic machine learning method (Kornejady et al., 2017) which was mostly applied, for species distribution, and was introduced by Phillips et al., 2004. The MaxEnt model is a Bayesian approach which focuses on maximizing probabilities, and it calculates the least biased estimates of the statistical probability, based on the input environmental factors (Phillips et al., 2006).

In this work, we used the maximum entropy model, taking into account seven landslide causing factors. The study area is represented by the South-

ern part of the city of Cluj–Napoca, considered as having the highest predisposition to landslides.

This study area was chosen for investigation due to its specific characteristics: the area is positioned at the periphery of the city and it is the most dynamic territory from the point of view of constructions, a fact determined by the rapid population growth. Rapid population growth generates an increase in needs, which is reflected in the number of new buildings. These districts are the most predisposed to landslides, but the anthropic factors have worsened the situation.

Among the urbanistic plans of the city of Cluj–Napoca, an official landslide susceptibility map was found (Fig.1). We have used the existing official map to confront it with our results obtained within this research. Unfortunately, comparing these two data, we could conclude the fact that, in the case of Cluj–Napoca, it becomes urgent to build a new landslide susceptibility map, because the existing data is not accurate and it is overdue. Therefore, the main aim of this study is to prepare an accurate and valid susceptibility map which can be useful for hazard mitigation and land–use/land planning strategies.

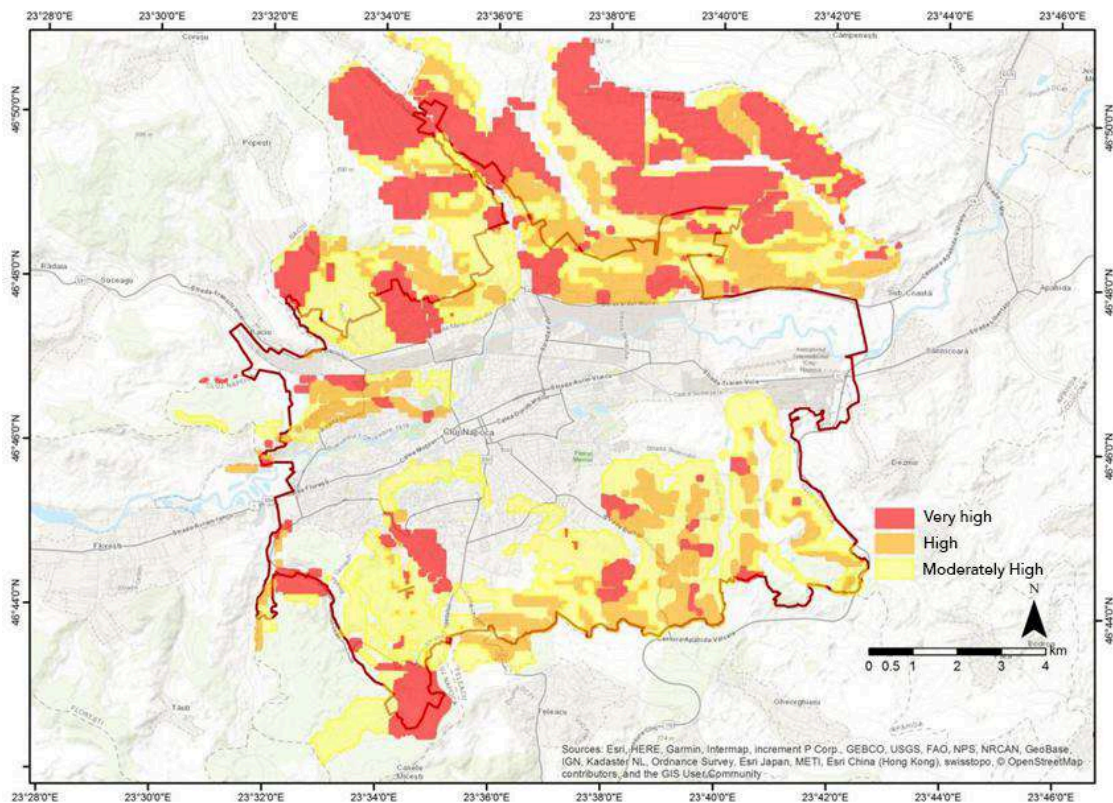


Figure 1 The official landslide susceptibility map (source: www.primariaclujnapoca.ro/proiectpug.html)

2. Study Area

The city of Cluj–Napoca is situated right at the connection spot between the Apuseni Mountains and the Transylvanian Plateau, at the confluence between the Someșul Mic and the Nadăș Rivers; it marks as a hilly region, thus representing a transition from the Gilău Mountains to the Someșan Plateau and the Transylvanian Plain.

Analysing the city map of Cluj–Napoca, it can be observed that many constructions, which were built until the XXth century, are based on the stable floodplain of the Someș River. Due to the rapid growth of the population, the districts of the city expanded rapidly towards the steeper slopes, favoured by the improvements in technology. This expansion trend contributes to increasing the land–

slide potential in these districts. The same process appears in the watershed of the Becaş stream.

The area under investigation is situated in the Southern part of the city of Cluj-Napoca, occupying a small part of the valley of the Someşul Mic River, situated on the Northern side of the Feleac Hills and was previously studied by Poszet, 2017 (Fig. 2). This territory (~25 km²) represents the watershed of the Becaş stream and it is the most exposed territory to landslides due to high anthropic pressure.

From a geological point of view, within the study area, numerous formations can be found, such as Paleogene (Upper Eocen and Early Oligocen), Neogene (Middle Miocene) and Upper Pleisto-

Within the study area, the following formations

cene formations, which are interleaved and form a complex geology; the surface of the entire area (including the terraces) is covered with slope deposits of varying thickness (Wanek and Poszet, 2011) (Fig.3).

The Western part of the study area is characterized by monocline structures and the Eastern part consists in diaper foldings.

In the Southern part of the area under investigation, Badenian formations appear under the Sarmatian ones and under a superficial layer of deposits and terraces. The youngest Quaternary deposits (Pleistocene and Holocene) create periglacial formations.

sion of sand, sandy-clay and sandstone; Dej Tuff

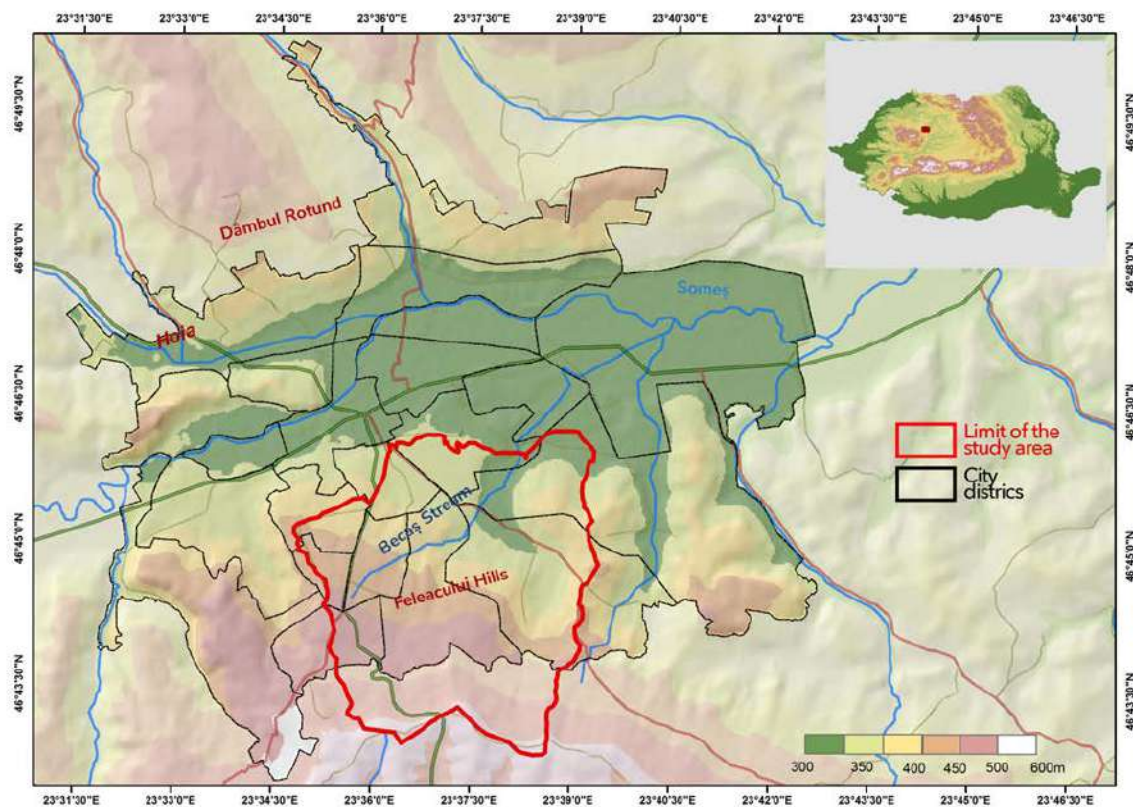


Figure 2 Geographical position of the study area within the Cluj-Napoca municipality

can be distinguished (Poszet, 2017, Fig. 3): Valea Nadăş Formation – (from the Upper Eocen) has a thickness of approximately 60–80 m and it consists in red clay, silt and sand. These characteristics induce the occurrence of landslides; Gruia Sandstone Formation – (from the Early Oligocen) has a thickness of 20–25 m and it is characterized by a succes-

Formation (from the Badenian) – has two levels of depth; the primary level contains marl and clay with a thickness of 20 m on which the second level was deposited (this contains tuff), with a thickness of 60 m. This formation is stable in terms of landslides; Iris Formation (from the Lower Sarmatian) – has various clasts deposits and interleaved tuff and sand. This

formation appears above the Badenian formations until the upper third of the northern slope of the Feleac Hills; Feleac Sandstone Formation – (from the Upper Sarmatian) Iris Formation was deposited on top of it; there is a juxtaposition of sand, sandstone and clay. These Sarmatian formations are very susceptible for landslides. In the Quaternary (the Holocene), the seven levels of the Someş River terraces took shape; these are stable in terms of landslides.

The morphostructure, morphometry and topography of the area under investigation are predisposing factors to landslide reactivation. Translational slides (type 12 according to Hungr et al., 2014) appear on the steep slopes of the southern part of the study area (>200); on the crowns and main scarps of the Pleistocene slidings rotational landslides can be observed (type 11).

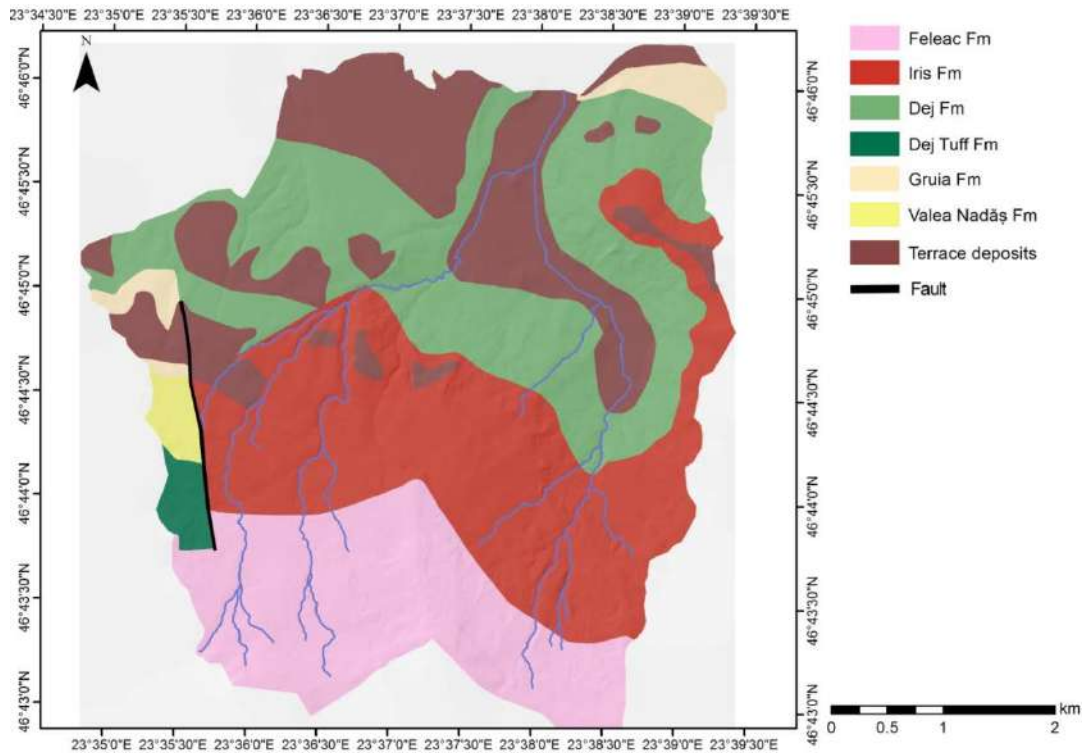


Figure 3 Distribution of the geological formations in the study area

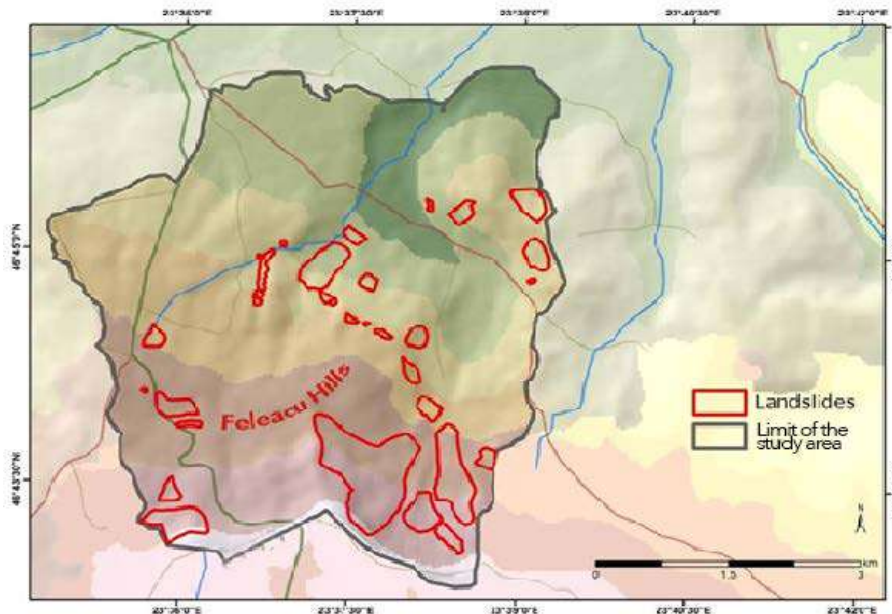


Figure 4 The distribution of the inventoried landslides in the study

The morphology has a high complexity and it is determined by geological, anthropic, climatic characteristics and fluvial erosion. Landslide features (crowns and main scarps) dating from the Pleistocene may be observed (Wanek and Poszet, 2011, Poszet, 2017). Over time, the mass movements were redeposited and temporary stabilized, but due to anthropic factors and failures, on top of the old Pleistocene landslides, many new slope failures have occurred or reactivated due to human presence.

The climate can be considered as a control factor in the present and the past geomorphological processes. In the Pleistocene, the major relief formations were created, on which the actual geomorphological processes were grafted (Poszet, 2017). The climate determines the rhythm, the length and the intensity of the actual geomorphological processes. The actual climate, which is dryer than in the previously mentioned periods, helped the stabilization process of mass movements, but the landslides can be easily reactivated by different factors.

The precipitation is an important landslide triggering factor, because mass movements are favored by long-lasting rainy periods, freeze-thaw and wetting-drying processes (Conforti et al. 2011; Kerekes et al., 2016).

3. Materials and methods

3.1. Materials

In this research, we used a wide range of data: topographical maps (1:5000, 1975), geological maps (1:5000, Wanek and Poszet, 2011), aerial imagery (ANCPI – National Institute of Cadaster (2012) and Google Earth 3D (2014)) and open-source data. These data were processed in ArcGIS software. In order to obtain a precise digital elevation model (DEM), we digitized the contour lines from the topographical maps. Interpolating the resulted vector, we generated a 1 x 1 m resolution DEM.

Based on the above mentioned data, we could identify and map all the landslides of the area, obtaining a landslide inventory map (Fig.4.) which was verified during field-works. The landslide inventory contains 34 landslides, with a total area of 25.8 km²,

which means that approximately 10% of the territory is covered by mass-movements.

After analyzing the landslide inventory that we compiled, we selected the landslide predisposing factors to be evaluated with our approach: lithology, topographical parameters (slope aspect, slope angle, depth of fragmentation, plan- and profile curvature) and land use.

The geomorphometric parameters were derived from the 1 x 1 m elevation model. To store all the information in a uniform thematic database, the pixel size for all products was set to 1 m:

i) *Slope aspect* (Fig. 5.a): the slope aspect influences landslide processes by producing major differences in climatic parameters, controlling the duration and angle of sunlight, evapotranspiration, moisture retention and vegetation cover type (Dai et al., 2001; Sidle and Ochiai; 2006; Rahmati et al., 2016; Kerekes et al., 2016);

ii) *Slope angle* (Fig. 5.b): in fact, this can be considered the most important causative factor, because all mass movements occur as a consequence of gravitational force (Liberatoscioli et al., 2017, Pradhan and Youssef, 2009). In our case, most of the mass movements occur near or inside the perimeter of the newly built areas, i.e. on slopes having values between 50 and 100;

iii) *Profile curvature and plan curvature* (Fig. 5.c and d): Profile curvature measures the rate of change of the slope; plan curvature influences the convergence or divergence of downhill flowing water. Therefore, both become important factors in landslide formation (Meinhardt et al., 2015);

iv) *Depth of fragmentation* (Fig. 5.e) is a well-known geomorphological indicator; it represents the relative altitudinal difference per square kilometer. This indicator shows an outgo rate of the relief's quantitative characteristics (Tatiana Constantinov et al., 2009).

Apart from the topographical parameters, another landslide predisposing factor that was considered in the analysis is geology. The type of mass-movements under investigation mostly occurs on sedimentary rocks like clays, marls or sandy-clay. The high occurrence of landslides in the study area is explained by the presence of Oligocene, Miocene

and Pleistocene sedimentary rocks, which cover the entire territory, presented in section 2.

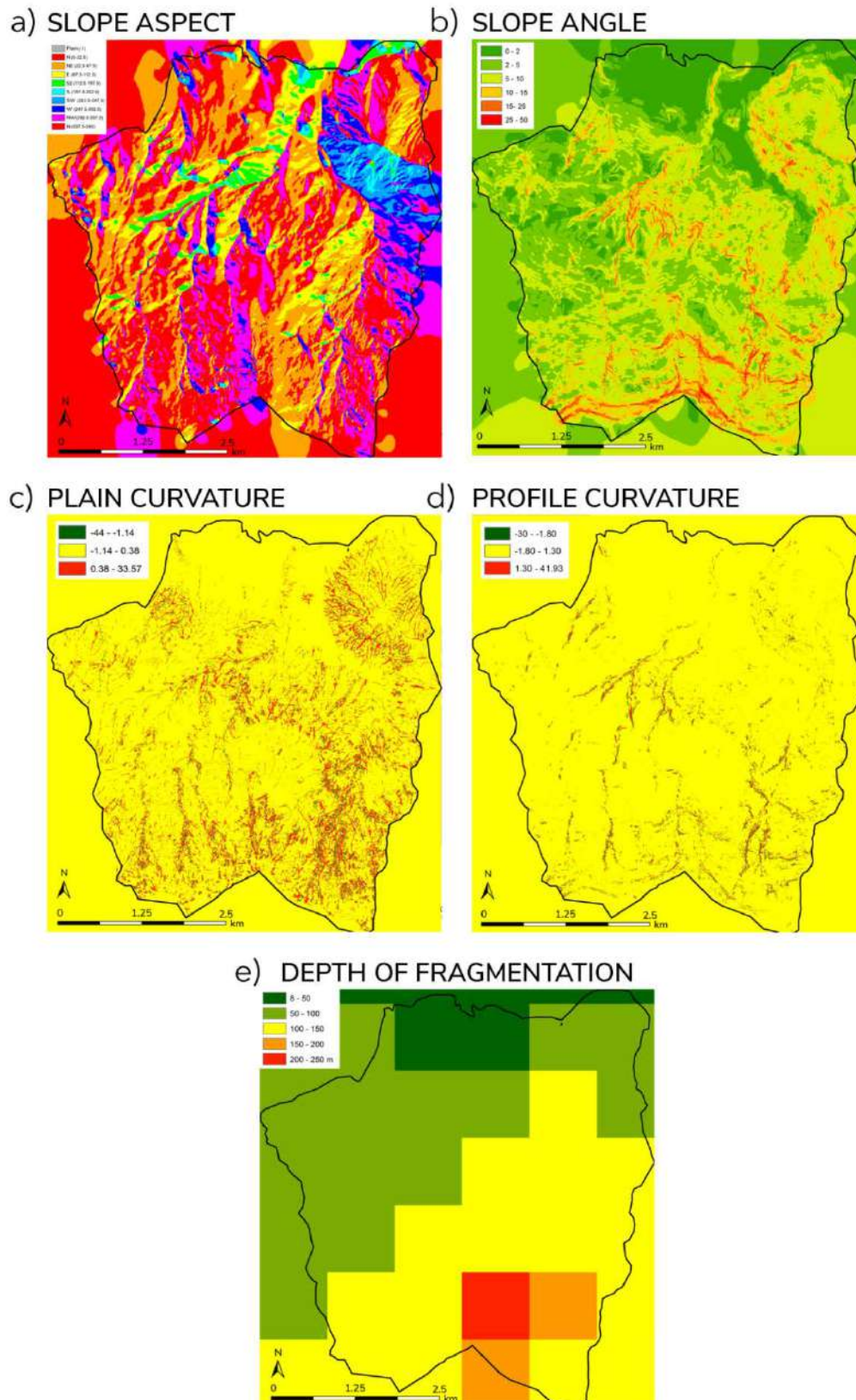


Figure 5 The morphometric parameters of the study area

Another landslide-causing factor is *land use* (Fig.6.) This vector layer was obtained from the open-source Corine Land Cover 2012 database.

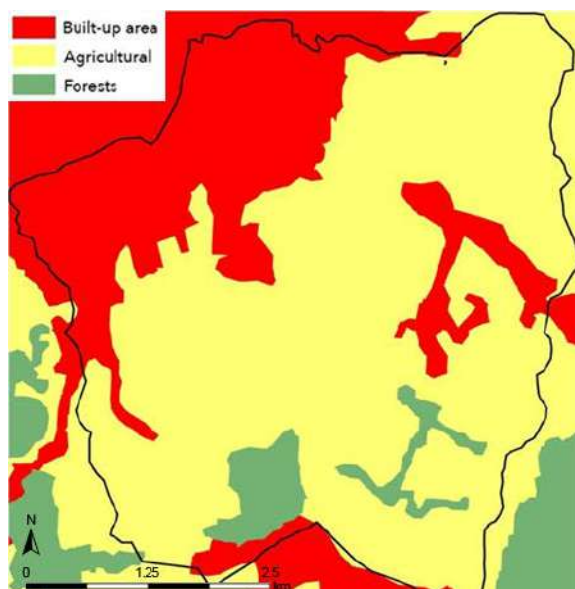


Figure 6 The land-use of the study area

3.2. Methods

The Maximum Entropy model is a statistical-probabilistic machine learning method, which has been applied to many different fields, such as environmental sciences for predicting species distribution, (being proposed for the first time by Phillips et al., 2004, for the case when just presence data is available in order to calculate the best positive distribution and used with a great accomplishment by Pearson et al., 2007, Elith et al., 2011; Bajat et al., 2011; Yang et al., 2013; Muscarella et al., 2014; Cao et al., 2016), groundwater mapping (Rahmati et al., 2016; Wahyudi et al., 2012), crop planting regionalization (Franklin, 2009; Yu et al., 2014; Jaime et al., 2015).

Recently, this method was applied with a great success in landslide susceptibility mapping: Liu et al., 2012, Reza et al., 2012; Vorpahl et al., 2012; Pourghasemi et al., 2012a; Felicísimo et al., 2013; Park, 2015; Davis and Blesius, 2015; Moosavi and Niazi, 2016; Lombardo et al., 2016; Kornejady et al., 2017. Pandey et al., 2018 investigated in India the territory along the state highway corridor Tipari to Ghuttu. This area has similar geological characteristics to our territory, characterized by complex geological formations, where clay is predominating. In

this study, the Maximum Entropy principle was applied with a high performance, indicating the fact that this model is suitable for territories with a complex geology.

The MaxEnt method has a complex math, but it does not require high precision surveys (Kornejady et al., 2017). Entropy is a measurement of disorder, imbalance and uncertainty of a system (Yufeng and Fengxiang, 2009). The Boltzmann principle describes the one-to-one relationship between the quantity of entropy of a system and the degree of disorder (Yufeng and Fengxiang, 2009).

The maximum entropy model originates from information theory and statistical mechanics. Based on Boltzmann's H-theorem, Shannon (1948) managed to define entropy, as the expected value (average) of the information contained in the existing data:

$$H(x) = E[I(X)] = E[-\ln(-\pi(x))] \quad (1)$$

where, E is the expected value operator and I is the information content.

Therefore, the expression can be defined as the negative natural logarithm of the probability distribution: the expected value of a random variable is the average of the experiment it represents (observed data) (Hamming, 1991; Ross, 2014). Jaynes (1957) determined the probability distribution generated from maximum entropy model as the least biased possible estimate on a given information, therefore, when using predictive algorithms, the most performant model should coincide with the one that has Maximum Entropy (Lombardo et al., 2016).

The MaxEnt model applied by Phillips et al., 2004, was conceived as a software by the Center for Biodiversity and Conservation at the American Museum of Natural History (AMNH) through a public-private partnership between the AMNH and AT&T-Research.

The advantages of the model in prediction analyses include the following: it requires only presence data using environmental information (factors) for the whole study area; it can utilize both continuous and categorical data; the MaxEnt probability distribution is amenable to analysis; MaxEnt is a generative approach, which means that it can be used

when we have limited amount of training data (landslides) (Phillips et al. 2004; Park, 2015).

As a prime step, the MaxEnt model calculates a uniform probability distribution function as a primary data. In this data, every single pixel is assumed to have the same probability of landslide occurrence (Kornejady et al., 2017). Following the previous step, some constraints modify this uniform probability distribution function to realize an optimal probability distribution of the phenomenon (landslides) over a spatial context (in our case: land use, geomorphological characteristics, geology) (Phillips et al., 2004; Phillips and Dudík, 2008; Kornejady et al., 2017; Elith et al., 2011). Different features are represented by different predisposing factors, but these features represent the corresponding functions of the predisposing factors (Phillips et al., 2006; Elith et al., 2011). Therefore, we can conclude that the features, factors, and constraints are closely connected to each other (Phillips et al., 2006; Elith et al., 2011, Halvoren, 2012).

According to Davis and Blesius, 2015, the presence-only characteristic of landslides makes the maximum entropy model, which is made for species habitat analysis, appealing; geomorphological events have many common characteristics with biological occurrences in that while they respond to environmental factors, mass movements can increase the likelihood of other events in the proximity of the affected area, due to morphological and hydrologic conditions.

A landslide represents a complex system characterized by the exchange of materials and energy with the environment, therefore, landslides can be measured using the information entropy model (Yang and Qiao, 2009). The entropy of a landslide represents the extent of the various factors which are controlling the development of a landslide (Pourghasemi et al., 2012b). The resulted entropy value can be used to calculate weights of the system (Yang et al., 2010).

In this work, the landslide susceptibility map was generated by the MaxEnt software (https://biodiversityinformatics.amnh.org/open_source/maxent/) using all the prepared thematic maps mentioned above.

After generating the landslide susceptibility map, map evaluation becomes necessary. In order to evaluate our susceptibility map, it is necessary to verify the results by using different statistical methods. Model evaluation is a fundamental step in comparing the predictions with real world data-sets, in order to evaluate their performance (Beguería, 2006).

The most used method is the receiver operating characteristic curve (ROC) method. The ROC method is a threshold independent method; it plots different accuracy values realized against the whole range of possible threshold values (Beguería, 2006; Deleo, 1993).

Therefore, applying the ROC curve method, we can assess the accuracy of the MaxEnt model by comparing the obtained susceptibility map (probability map) with the landslide inventory (Chen et al., 2016; Beguería, 2006; Petrea et al., 2014; Bayes, 2015).

The ROC curve was realized using the number of correctly (true-positive) and the number of incorrectly (false-positive) classified pixels (Kavzoglu et al., 2015). The values from the OX axis represent surfaces, with no landslides, where the applied model has calculated a high probability (false positive). The values from the OY axis coincide with the surfaces with landslides, where the model has determined a high probability (true positive) (Petrea et al., 2014).

4. Results and Discussions

After running the MaxEnt software, the resulted values were grouped into five classes (Fig. 7A.), using the natural breaks distribution: representing very low, low, moderate, high and very high susceptibility levels. To assess the model's performance, we used the AUC (Area Under the Curve) method.

The prediction accuracy is high when the values are higher than 0.9. AUC values between 0.7 and 0.9 represent reasonable discrimination ability; furthermore, values under 0.5 indicate a poor quality of the results (Swets, 1988). With the maximum entropy model, we obtained a 0.844 AUC value (Fig. 7B), which means that our results have reasonable discrimination ability, therefore can be used for further studies.

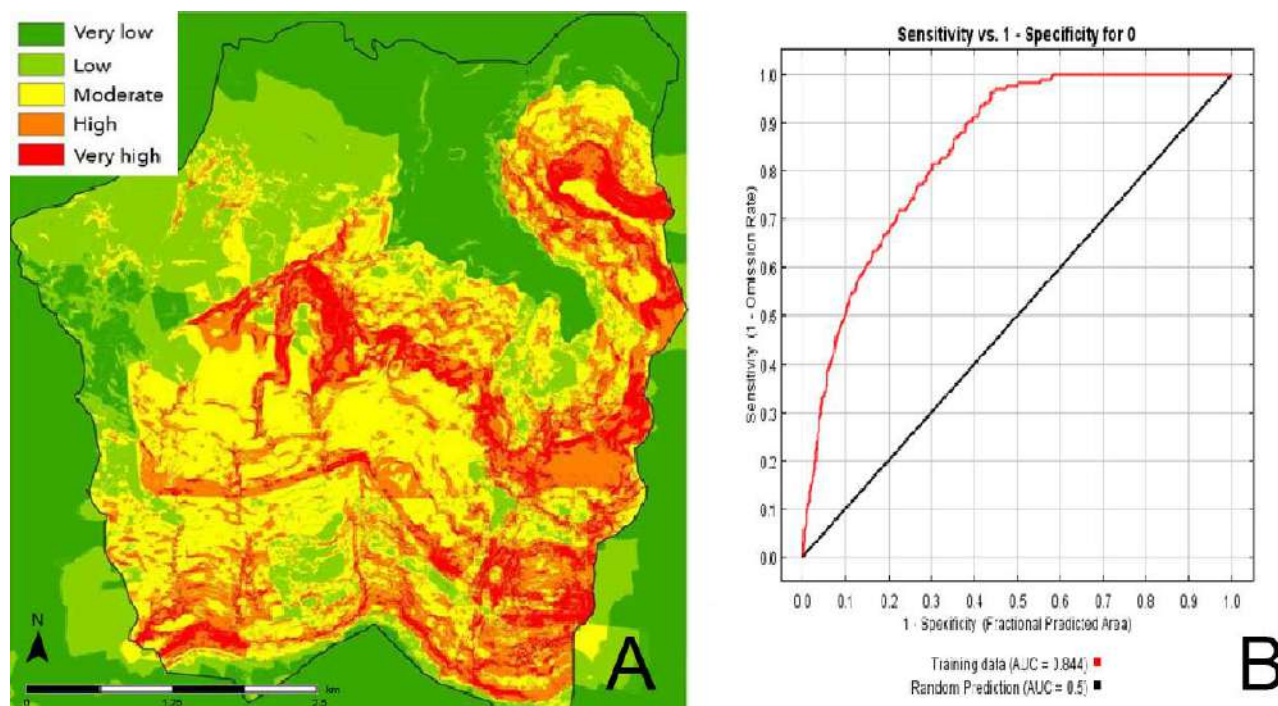


Figure 7 The resulted landslide susceptibility map for the study area (A) and the ROC curve of the susceptibility map (B)

On this map we can observe the fact that the highly prone areas outlined by the MaxEnt method are representing the Pleistocene landslides.

Analysing the resulted map (Fig. 7A), we obtained that 22% of the territory is included in the high and very high susceptibility category, being represented by steep slopes (most of them under constructions) and torrential formations. This category is mainly represented by the upper part of the slope of the study area, where we can find deep rotational Pleistocene landslides. On the surface of these deep landslides we can find the newly formed shallow ones. Due to the location of these landslides, we can say that scarp areas with high susceptibility are the most prone to reactivation. 21% of the area is moderately predisposed to mass-movements; this category is mainly represented by the middle part of the slope, where shallow translational landslides are present. 57% of the area falls into the low and very low susceptibility category, which is represented by the flat areas of the Someş floodplain. These low values are related to anthropic features like road and building embankments.

Studying the landslide inventory, 56.34% of the landslide pixels are included in the high and very

high category; 34.52% in the moderate category and 9.12% are included in the low and very low susceptibility class. These results also prove a high reliability of the model applied on the study area. Thus, we can conclude that the most stable landforms are the terraces of the rivers and the floodplain of the Someş River.

We can observe from the resulted map and from numerous field-trips the fact that in the southern part of the study area there were many abandoned orchards and meadows, near them deforestations were made. Nowadays, these areas gradually become built-up areas, therefore accentuating further mass-movements. Sparsely vegetated areas are more exposed to denudation processes, because the roots of the vegetation have a soil-binding power (Bayes, 2015, Kerekes et al., 2016). Between 2005 and 2008, every single safe zone in terms of landslides was shadowed by built-up areas; therefore, beginning with 2008 the constructions are spreading among the most landslide susceptible areas. In only 10 years, the surface of the built-up areas has increased with 69% compared to the total built-up area surface in 2005 (Fig. 8).

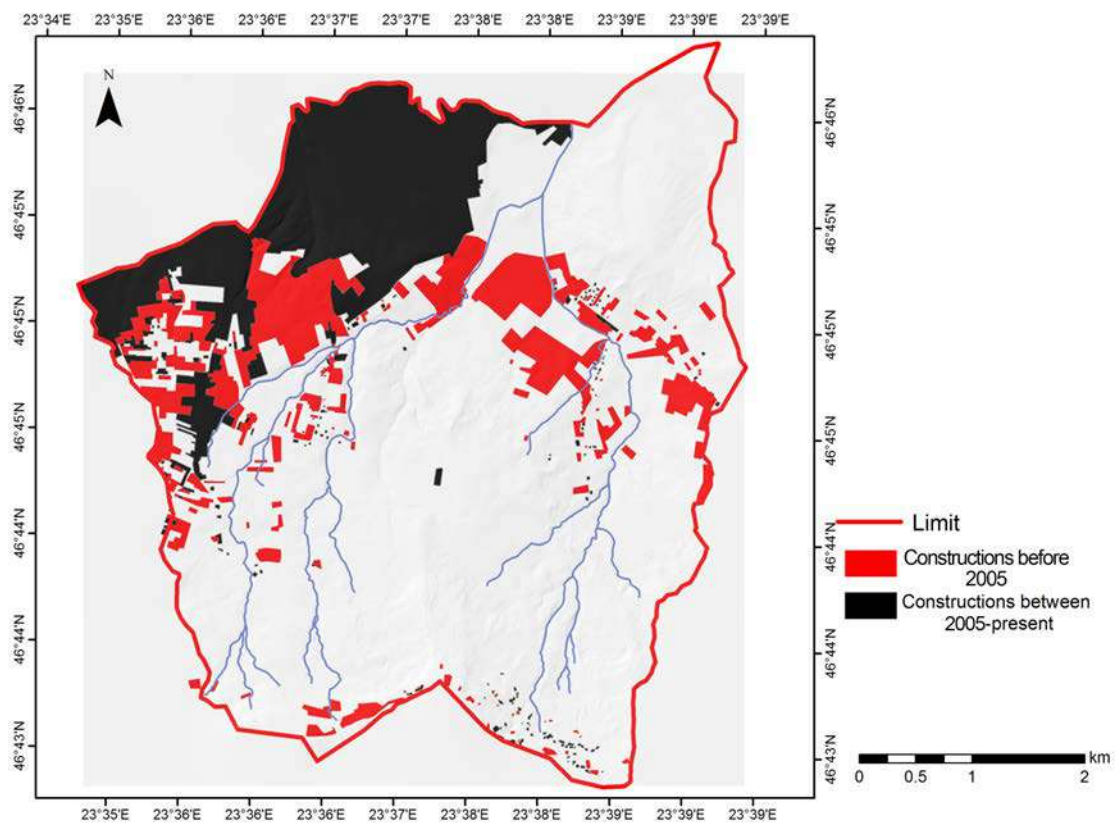


Figure 8 The built-up areas of the investigated area

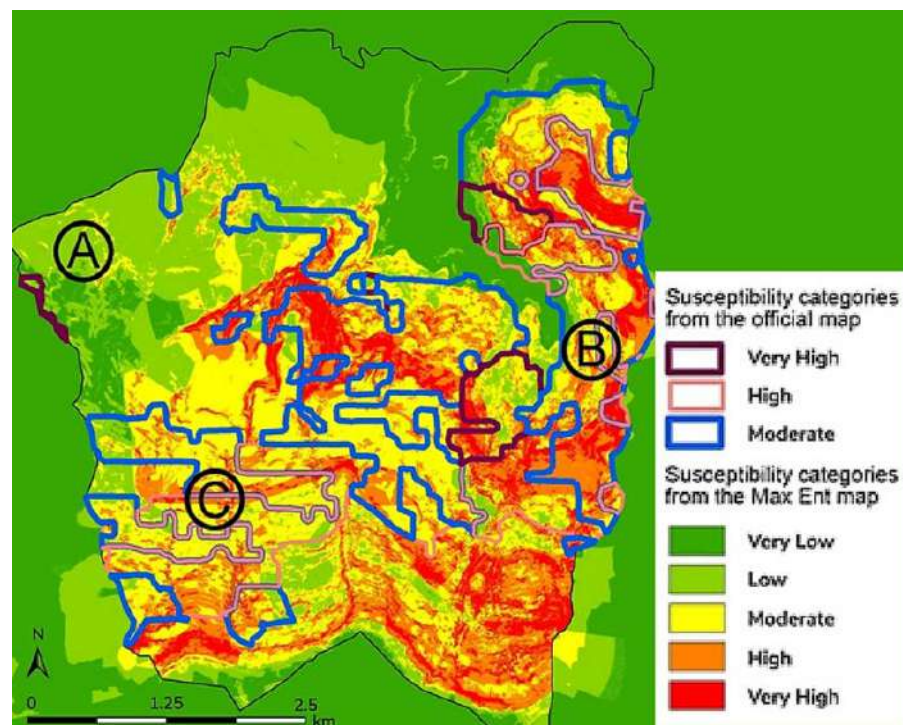


Figure 9 Map displaying the results of confronting the MaxEnt susceptibility map to the official landslide susceptibility map of the study area



Figure 10 Landslides caused by constructions (A) and affected the roads (B, C)

Thus, the main reason why we have chosen this study area is the fact that the extent of the built-up areas is still modest and there are many freely observable areas. Because of the high speed of constructions, the private areas will prevent further studies. Therefore, it becomes necessary to analyse this territory to raise awareness about the existing situation, until it is not too late. The constructions are relatively new, therefore primary mass-movements appear on the roads near the built-up areas (Fig. 9 and Fig.10 A, B, C). The number of constructions implies the increase of the road traffic therefore the trepidation causes landslides (Fig.10 B, C).

In order to calculate the factor contribution, the MaxEnt software used the jackknife technique, which gives the relative contributions of the environmental variables (Park, 2015). This represents the percent of a specific factor being used in the modelling process; therefore these factors have useful

information compared to the rest (Kornejady et al., 2017). As a result, we can see in Table 1 the fact that the lithology, slope and human activity have a significant role in landslide occurrence.

Table 1 Percent contribution of the landslide causing factors

Variable	Percent contribution
Lithology	56.9
Slope Angle	34.2
Land-use	7.8
Profile curvature	0.5
Depth of fragmentation	0.3
Plan curvature	0.1
Slope Aspect	0.1

Therefore, the results indicate the fact that landslides could be triggered by natural causes like geology and slope angle, as well as human activity (deforestation, constructions).

Comparing our results with the existing official landslide susceptibility map (Fig. 9) we obtained that the susceptibility categories of the official map do not coincide with ours (the official map does not show the low and very low susceptibility levels). We used the overlay method (on a pixel level), and found that, in case of the official map, the most susceptible areas are not properly mapped, therefore a lot of data is missing (we can observe a 21% matching from the total area between the two maps). Based on the official map, we could calculate the fact that 0.46% of our landslide inventory is situated in the very high, 1.06% in the high and 0.6%, in the moderate susceptibility category of the official map. For the rest of the content, there isn't any data available therefore we cannot render a match. Furthermore, we don't know the method which was used to generate the official map, just the results that can be seen. The high degree of the discrepancy is caused by the difference in the resolution of the landslide mapping and susceptibility map generation. We can deduce that the resolution of the official map is approximately 100 m. The map generated by us has 1 m resolution, thus we could identify more landslides, improving the accuracy of the analysis.

5. Conclusion

This study reconfirms that generating a valid landslide susceptibility map is a complex issue: we must put effort in compiling a landslide database as complete and as accurate as possible for the area under investigation; it is necessary to make a complex analysis of the existing statistical models in order to choose the most appropriate to be applied on the study area.

In this study, maximum entropy model was successfully used and tested on the Southern part of the city of Cluj–Napoca (Becaş stream watershed), yielding a high AUC rate (0.844), resulted from the generated ROC curve. We identified (based on long-term observations, many field–trips and studies)

seven landslide causing factors: slope, aspect, profile and plan curvature, geology and land–use.

As a result, we obtained the fact that the most prone territories are the steep slopes of the Pleistocene landslides where we can observe intense human activity; 56.34% of the slides occur on these territories, thus we can conclude that the most severe predisposing factors are the human induced activities. The floodplain of the Someş River and the gentle dip–slopes are characterized by low and very low susceptibility.

Future work will focus on a landslide susceptibility map for the entire Cluj–Napoca municipality because in the recent period, due to anthropic factors, many landslides have been reactivated in other important sectors of the city (Fig. 11) has been destroyed due to thoughtless construction works.



Figure 11 Landslide in a recreational park, 2016

Therefore, prevention of such mass movements becomes possible just with a valid susceptibility map and responsible human activity.

Current jurisdiction specifies that the drillings made by geotechnicians should have just 4 m depth, therefore these drillings don't reach the base of the slided blocks; few drillings are deeper than 4 m, which is a huge problem, because the Pleistocene mass movements are profound (Tövissiné Losonczy Ibolya, 2004) and many buildings are constructed on them. As an efficient prevention method, geotechnicians should make deeper drillings in order to precisely see the structure, texture, cohesion and composition of the deposits. Also,

every building foundation should be reinforced with a good amount of steel and concrete. A problem with this method is that it can cause overloading. Nonetheless, people must be informed by local authorities of the existing situation and the risks they are facing.

Future work is needed to improve the resulted map in order to generate future disaster mitigation works and land planning strategies. Further case studies should be performed using a greater number of data sets and landslide causing factors in order to improve the susceptibility map.

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Quantifying water erosion hazard in central Tunisia A case study: the Leben watershed

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ABSTRACT

This study was performed in order to quantify soil erosion in the arid lands of central Tunisia. To do this we used geographic information systems, remote sensing and the RUSLE prediction model. While the study area is characterized by reduced rainfall quantities and therefore a low *R* factor, other factors such as high soil erodibility, low vegetation cover and steep slopes resulted in locally high erosion rates. Indeed, the calculation of RUSLE factors revealed values between 0 and 731 $\text{tones} \times \text{ha}^{-1} \times \text{year}^{-1}$ with an average of 3.84 $\text{tones} \times \text{ha}^{-1} \times \text{year}^{-1}$. Whereas most catchment lands have low levels, the relief surrounding the catchment has relatively high rates due to the high value of slopes, the skeletal soils that compose them, but also due to the absence of protection, reforestation or crop management. This study also showed that despite their small extension compared to the watershed, main streams sectors, and especially their banks, record the highest erosion rates mainly due to the flow accumulation in the downstream sectors. Relative protection of the piedmont, which has saved them from higher erosion, can only be considered efficient and sustainable if the upstream hillsides are also protected. The comparison with previous studies carried out in Tunisia shows similarities in the results of some of them, considering the natural geographical variation of the *R* factor.

KEYWORDS

Water erosion; RUSLE; Tunisia; arid lands

1. Introduction

Soil water erosion is one of the main threats to arable lands sustainability. Tunisia is one of the Mediterranean countries most affected by this hazard. Combating land erosion in Tunisia is an old phe-

nomenon and has become over time more widespread and more technical. Although studies on soil erosion have multiplied over time, the arid regions of Tunisia have not, to our knowledge, been of particular interest for this type of study. This is probably

due to low precipitation that suggests low erosion, which this study is trying to contradict. Indeed, the scarcity of rainfall does not mean that central Tunisia lands are not affected by water erosion. On the contrary, precipitations in these regions follow wet cycles and less wet or even dry cycles. During wet periods, rainfalls often cause flooding with runoff and erosion values comparable to the more humid regions of northern Tunisia. Studies to quantify soil erosion began a few decades ago, but spatialization and mapping techniques were unreliable and resource-intensive. Today, this work is made easier by tools such as geographic information systems, remote sensing and digital models that can be integrated into them. This study intends to use these tools to attempt to quantify erosion based on the RUSLE model. This model is established on experimental studies plots and it is in continuous improvement (Wischmeier et al., 1965; Renard et al., 1997). The objective of such a model is not to stop erosion but to know the high-risk areas and then to

act on the controllable factors such as slope length (shortening it by contouring or terracing), or by well choosing crops and protection types (Masson, 1971).

2. Study area

The study area, with an area of 1470 km², is located in central Tunisia between 9°10' to 9°50'E and 34°29' to 34°51'N (Fig. 1). Altitudes vary from 115 m in the central plains to 986 m in the relief bordering the watershed. The Meloussi and Majoura mountains border the catchment on the north and west sides, the South Bouhedma Mountain and the Mehiri-Zebbeus mountain on the east side (Fig. 2). The major part of the basin corresponds on the contrary to vast plains which are disturbed only locally by Kodiât Ouled Amara plateau. These reliefs generally correspond to anticlines bordering two large synclines, partially separated to the west by a small anticline convexity.

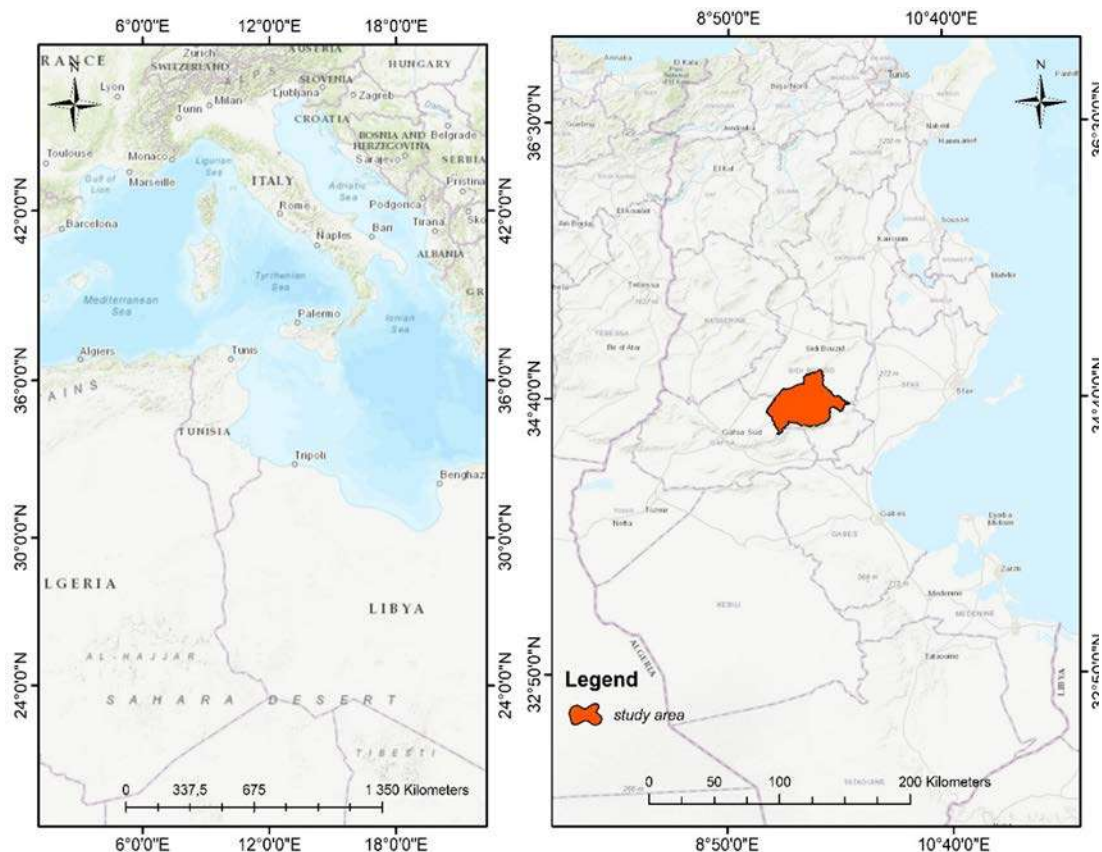


Figure 1 Location of the study area

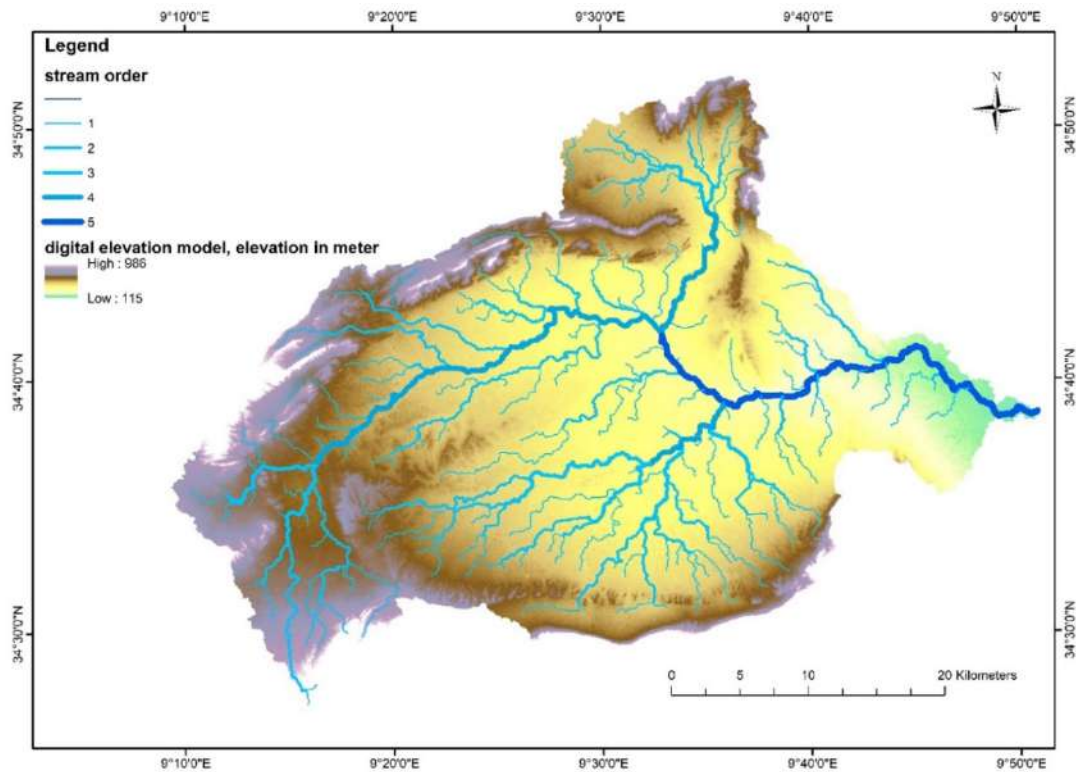


Figure 2 Digital elevation model and stream

The outcropping rocks on the reliefs are essentially Cretaceous limestone and dolomite except the reliefs of the east where they are composed by gypsum and clay–gypsum layers. Clay outcrops are particularly present in the south–eastern part of Dribika and Nadour of Triassic constitution. Neogene deposits (clay, conglomerate, sand) often occupy the northern and eastern piedmonts with gypsum or calcareous crusts. Inherited or actual sand accumulations litter the relief of the southern limit of the catchment. Most of the central plains are alluvial layers from the Upper Pleistocene and Holocene. The study area belongs mainly to the upper arid climate stage, a small part to the south–east to the cool lower arid and a band corresponding to the Majoura Mountain (north–west of the basin) belongs to the temperate lower semi–arid stage. The relief surrounding the watershed is covered by a degraded scrubland of *Phoenicia juniper*. While the western piedmonts are covered by an esparto steppe, the southern piedmont is the site of a sandy jujube steppe. The eastern and north–eastern gypsum lands are covered with halophile vegetation.

3. Materials and methods

Presentation of RUSLE model: This model consists of five independent parameters which, multiplied according to the following equation result in the final soil loss rate (Eq. 1) (Fig. 3).

$$A = R \times K \times C \times P \times L \times S \quad (1)$$

where, A = annual rate of soil loss in $\text{tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$, R = rainfall erosivity factor, K = soil erodibility factor, C = crop cover factor, P = practice support factor, LS = topography factor.

The rainfall erosivity factor R , as defined by Wischmeier and Smith (1978) (Eq. 2), means the rain kinetic energy E multiplied by the maximum rainfall intensity (I_{30}) for 30 min. This index represents the potential erosive risks in a specific region. In our study, given the impossibility to apply the Wischmeier and Smith equation, we opted for the formula developed by Renard and Freimund (1994) based on the modified Fournier index (MFI) which establishes the relationship between the MFI and the R factor with a high determination coefficient ($r^2=81$) (Eq. 3a) (Renard and Freimund, 1994). The MFI reflects both annual and monthly average and

will avoid overestimating of R factor values in a region characterized principally by its aridity (Eq. 4) (Arnoldus, 1980). The MFI and R factor were calculated on the rainfall data of 7 stations located around but also inside the catchment area, for the period from 2002 to 2011. This period was chosen because it has no data gaps.

$$R = (E_c) \times (I_{30}) \quad (\text{Wischmeier and Smith, 1978}) \quad (2)$$

where, R is the erosivity factor for an event in $\text{MJ} \times \text{mm} \times \text{ha}^{-1} \times \text{hr}^{-1} \times \text{year}^{-1}$, E_c is the total rain kinetic energy in MJ/h and I_{30} is the maximum rain intensity in 30 minutes in mm/h .

$$R = 0.07397 \times F^{1.847} \text{ when } F < 55 \text{ mm} \quad (3)(a)$$

$$R = 95.77 - 6.081F + 0.477 \times F^2 \text{ when } F > 55 \text{ mm} \quad (b)$$

where F is the Modified Fournier Index

$$F = \sum_{i=1}^{12} \frac{(P_i)^2}{P} \quad (4)$$

where P_i is the monthly rainfall average and P is the annual average.

The soil erodibility factor K is a function of organic matter, texture, permeability and soil profile structure. The initial equation of Wischmeier et al., 1978 (Eq. 5) has often been used to define tables containing soil k values. The higher this index, the higher the soil erosive potential. In our study we extracted the soil map from the *Carte Agricole* (agricultural map, digital database) of the ministry of agriculture. Then we assigned the k values available in literature from Tunisia or elsewhere (Ben Cheikha et al., 2008; Andersson, 2010; Zante, 2003; Kort, 2013; Leguedois, 2003).

$$K = A \times B \times C \times D \times 0.1317 \quad (5)$$

where:

$$A = [0.2 \times 0.3 \exp(0.0256 \text{ sand}(1 - \text{silt}/100))]^{0.2}$$

$$B = \left[\frac{\text{silt}}{\text{clay} + \text{silt}} \right]^{0.2}$$

$$C = \left[1.0 - \frac{0.25 \text{ clay}}{(\text{clay} + \exp[(3.72 - 2.95 \text{ clay}]))} \right]$$

$$D = \left[1.0 - \frac{0.7 \text{ sand } 1}{(\text{sand } 1 + \exp[(-5.41 + 22.9 \text{ sand } 1)])} \right]$$

The topographic factor LS is a function of both slope length L and slope steepness S . It means the slope contribution to increase erosion potential in the study area. To obtain LS value map we applied the following equations: McCool et al., 1987; Des-

met et al., 1996 (Eq. 6). We used the 28 m resolution Astgdem digital elevation model. This choice was dictated by the relatively large area of the catchment but also the absence of a more detailed DEM covering the basin.

$$LS = L \times S \quad (6)$$

$$L = \left(\frac{\lambda}{22.13} \right)^m$$

$$m = \frac{F}{1+F}$$

$$F = \frac{(\sin \beta \times 0.01745/0.0896)}{3(\sin \beta \times 0.01745)^{0.8} + 0.56}$$

m = index of slope's length factor

$$L_{(i,j)} = \frac{[A_{(i,j)} + D^2]^{m+1} - A_{(i,j)}^{m+1}}{X^m D^{m+2} (22.13)^m}$$

where: β = Slope layer in ArcGis (in degrees), A = Flow accumulation layer (ArcGis), D = grid cell size (meters), X = coefficient that corrects the length of flow way through a raster cell.

S -factor

$$\text{when } \tan \beta_{(i,j)} > 0.09 \quad S_{(i,j)} = 10.8 \sin \beta_{(i,j)} + 0.03$$

$$\text{when } \tan \beta_{(i,j)} \geq 0.09 \quad S_{(i,j)} = 16.8 \sin \beta_{(i,j)} - 0.5$$

Crop management factor C refers to the role of vegetation in protecting soils from erosion. It includes both natural and cropland vegetation. Today, remote sensing makes it much easier to study crop cover. In our study we used a Sentinel_2_1C image with a resolution of 10 m. Image processing and field confirmation allowed us to make a supervised classification. This has permitted us to have land use classes to which we have assigned the C factor values from literature (Masson, 1971).

The conservation practice factor P considers man-made facilities and agricultural practices (ploughing, crop rotation, etc.) used to reduce runoff and erosion. This factor varies from total protection (0) to no protection (1). In our study we firstly digitized conservation practices from Google Earth. Then we exported polygons to ArcGis where we integrated them into the class slope map in percentages. Afterwards we extracted and pasted them with the untreated area and assigned P values to these polygons according to the slope classes. The untreated area was given a value of 1.

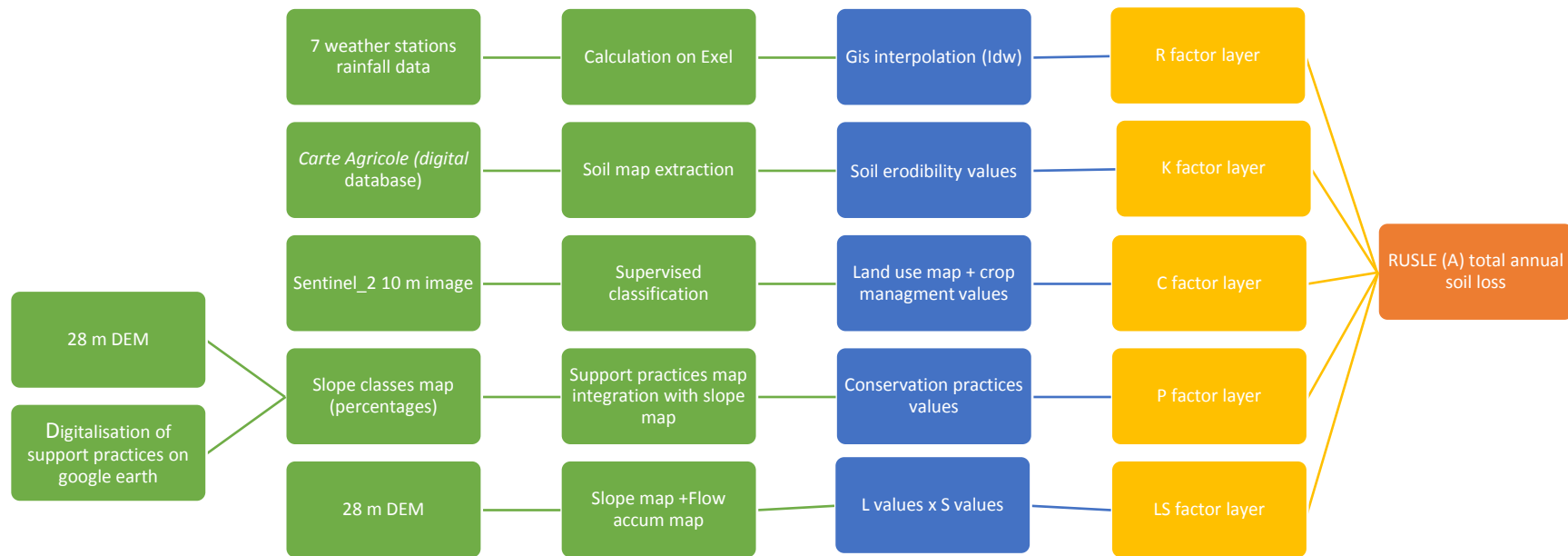


Figure 3 Methodology flow chart

4. Results and discussion

After the data collection phase, we proceeded to calculate model factors, one by one, then to multiply them in the math algebra tool of the ArcGis software. The results are as follows:

– Rainfall erosivity factor R : after interpolation with the Idw tool and extraction in the ArcGis software, the study region shows an R factor evolving between 20.58 and 32.59 $\text{MJ}\times\text{mm}\times\text{ha}^{-1}\times\text{hr}^{-1}\times\text{year}^{-1}$ (Table 1) (Fig. 4). These values are very low com-

pared to other regions but are consistent with their climate and the amount of rainfall the region receives. Moreover, some studies cite R values comparable to these. They particularly are in concordance with the results of Zante et al., (2003). This author recorded for Abdessadok watershed (Tunisian dorsal, Siliana region, 345 mm) values of 64 $\text{MJ}\times\text{mm}\times\text{ha}^{-1}\times\text{hr}^{-1}\times\text{year}^{-1}$ as average with values of 25 for the dry year and 93 $\text{MJ}\times\text{mm}\times\text{ha}^{-1}\times\text{hr}^{-1}\times\text{yr}^{-1}$ for the very rainy year.

Table 1 Weather stations, annual rainfall average and R factor for 2002–2011 period

Stations	LONG	LAT	Rainfall Annual average (mm)	Modified Fournier index	R factor
Bir Ali	10.095789	34.740684	186.68	20.61	19.77
Skhira	10.0699	34.299882	187.87	22.12	22.53
Mezzouna	9.84223	34.580594	182.79	21.90	22.13
Meknassy	9.602237	34.604781	197.52	21.06	20.59
Bouzayane	9.427244	34.573043	210.85	24.11	26.42
Sidi Bouzid	9.483061	35.037113	308.11	33.17	47.63
Gafsa	8.784391	34.422414	193.6	14.53	10.37

– The soil erodibility factor K : this factor was determined first by extracting the soil map as polygon files from the Carte Agricole (national agricultural mapping program) and then by assigning K values from studies performed in Tunisia or elsewhere. According to experimental work by Dumas (1965), (Table 3), soils constituting the study area belong to soils with high erodibility and very high erodibility (between 0.04 and 0.075 $\text{tones}\times\text{ha}\times\text{h}/\text{MJ}\times\text{mm}\times\text{ha}$) (Fig. 5) (Table 2). Soils very sensitive to erosion are confined in the moun-

tain belt surrounding the catchment. These are mainly raw mineral soils and low developed soils. But the rest of the land in the watershed is also composed of highly erodible soils. The spatial distribution of the different soil types shows the diversity of the formations occupying the basin. Thus, low developed soils are the most widespread in the catchment (33%). Then we find complex soils with 15%, followed by isohumic soils with 18% of the watershed. Raw mineral soils and rendzine occupy respectively 13% and 8% of the total surface.

Table 2 Soil types, K factor values and sources

Soil type	K factor values	ha	%	Source
Rendzine	0.05	12394.2	8.42	Ben cheikha et al.(2008)
Raw mineral soils	0.075	19895.47	13.52	Ben cheikha et al.(2008)
Low developed soils	0.055	49720.27	33.80	Ben cheikha et al.(2008)
Brown calcareous soils	0.046	9699.305	6.59	Ben cheikha et al.(2008)
Iso humic brown soils	0.042	27289.38	18.55	Kort (2013)
Halomorphic soils	0.04	4620.358	3.14	Kort (2013)
Complex soils	0.05	22274.69	15.14	Andersson (2010)
Stream bed	0.046	1172.018	0.79	Zante (2003)

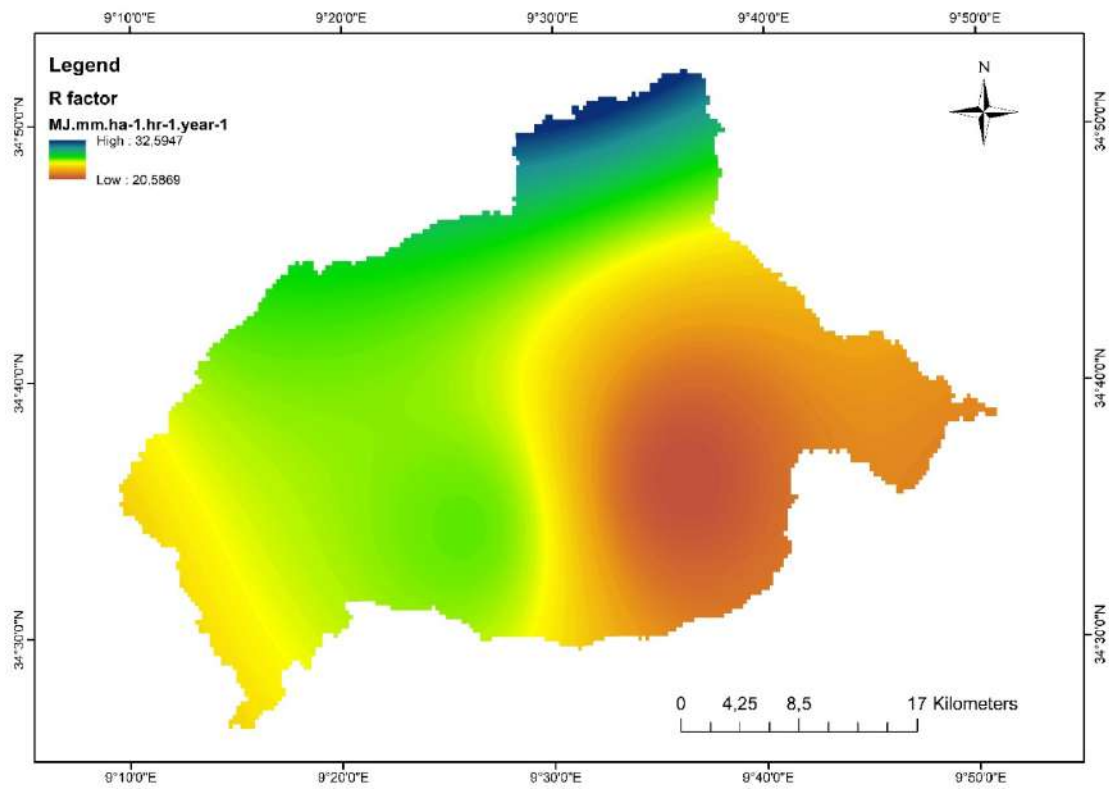


Figure 4 R factor

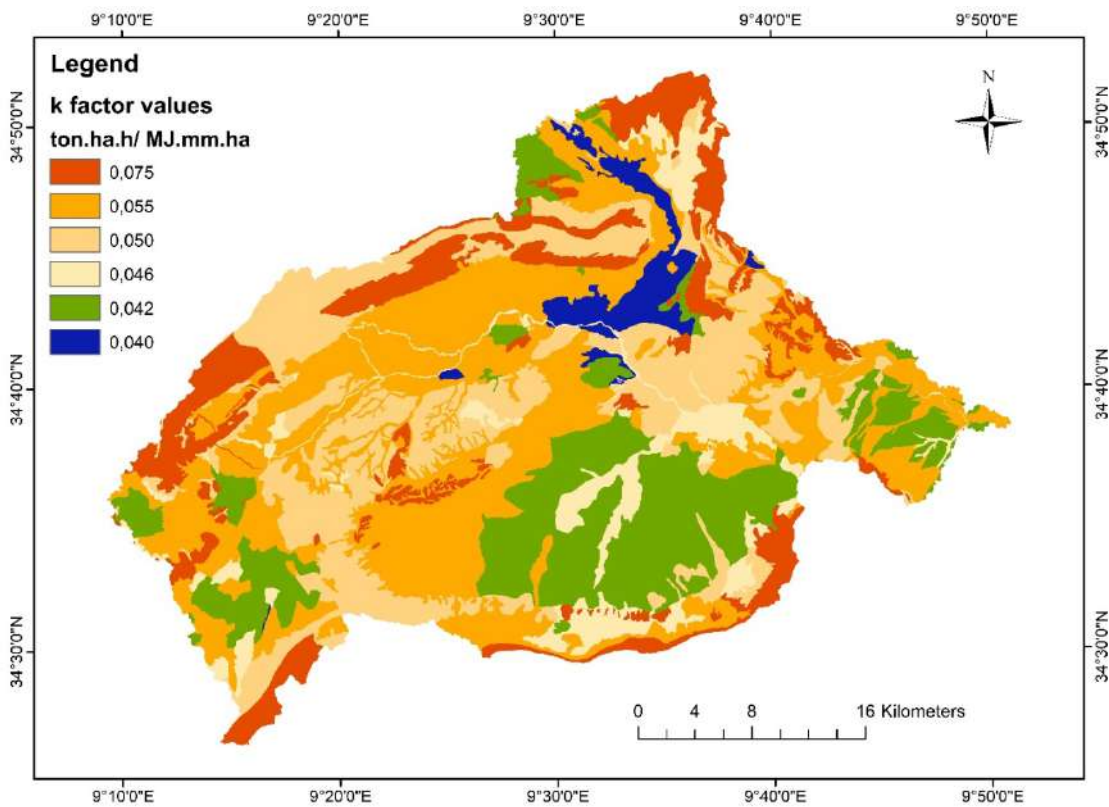


Figure 5 K factor

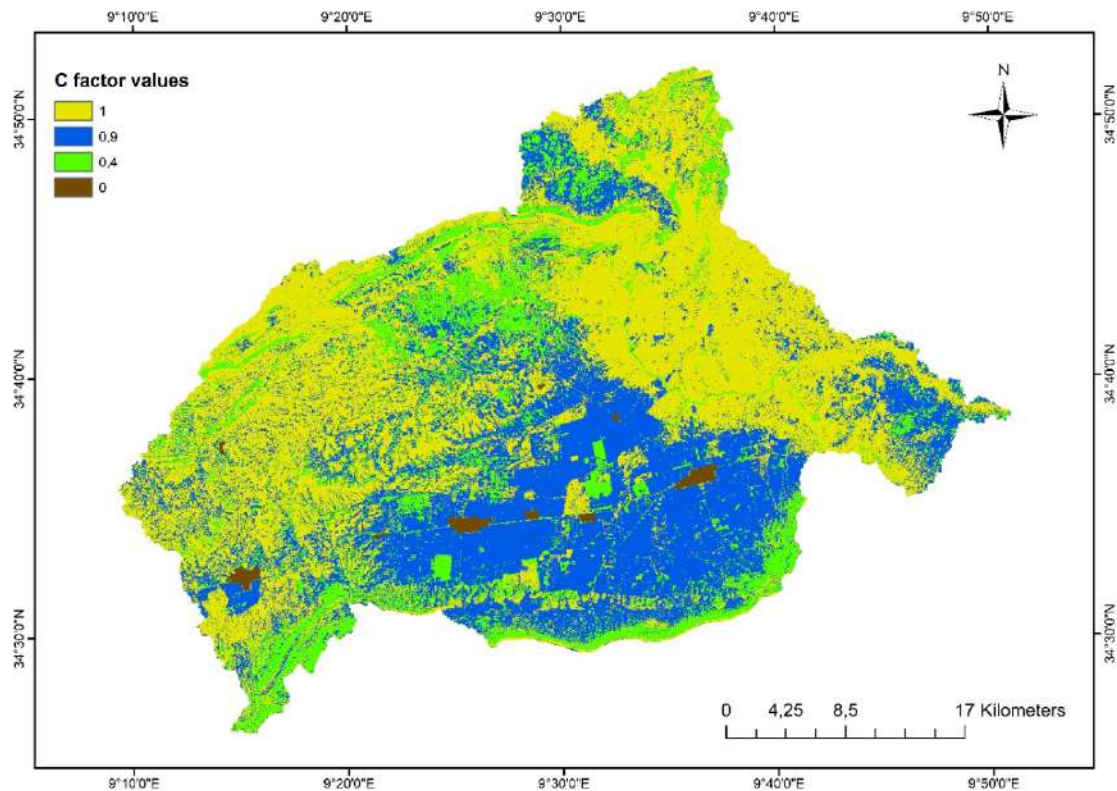


Figure 6 *C* factor

Table 3 Soil erodibility classes according to Dumas (1965) after conversion to SI units system (Foster et al., 1981)

Values classes	Erodibility classes
$K < 0.0065$	very lowly erodible soil
$0.0065 < K < 0.0131$	slightly erodible soil
$0.0131 < K < 0.0263$	moderately erodible soil
$0.0263 < K < 0.0526$	highly erodible soil
$0.0526 < K < 0.079$	very highly erodible soil

– Crop management factor *C*: this factor was first determined from a Sentinel 2 satellite image classification and the identification of the different homogeneous land use zones. Then the *C* values available in the literature were assigned to these areas (Masson, 1971) (Table 4). The following values are given: 0.9 for olive groves which occupy 35% of the catchment area, 0.4 for intensively cultivated areas, i.e. 15% of the catchment area, and 1 for bare soil which represents approximately half of the catchment area (Fig. 6) (Table 5).

– The conservation practices factor *P*: this factor was first created by digitizing the treated areas,

essentially by contour benches from Google Earth. Then, polygons were integrated with slope map classified in percentages. Afterwards, they were extracted from the unprotected area. While the maximum value 1 has been assigned to these, the values of the treated areas depend on their slope class. Results are as follow: untreated area represents the major part of the watershed: 89%. The remaining 10% has *P* values oscillating between 0.55 and 0.9 (Fig. 8) (Table 6).

Table 4 *C* factor table according to Masson (1971)

Vegetation type	<i>C</i> factor
Bare land, fallow land	1
Fruit arboriculture	0.9
Winter wheat	0.7
Cereal crop rotation	0.4
Forages	0.2
Cereal crop rotation + fodder	0.1 to 0.01
Improved pastures	0.01

Table 5 Crop management values and sources

Crop management	C factor values	ha	%	Source
Bare soils, bare rocks	1	71240.50154	48.44	Masson (1971)
Olive groves	0.9	51497.14704	35.01	Masson (1971)
Irrigated crops and intensive plots	0.4	23343.46525	15.87	Masson (1971)
Built-up areas	0	971.1252788	0.66	–

Table 6 P factor values and sources

Slope classes	P factor values (contouring)	ha	%	Source
0–7	0.55	9232.803	6.27	Shin (1999)
7–11.3	0.6	4069.721	2.76	Shin (1999)
11.3–17.6	0.8	2005.117	1.36	Shin (1999)
17.6–26.8	0.9	467.7729	0.31	Shin (1999)
>26.8	1	131296.4	89.27	Shin (1999)

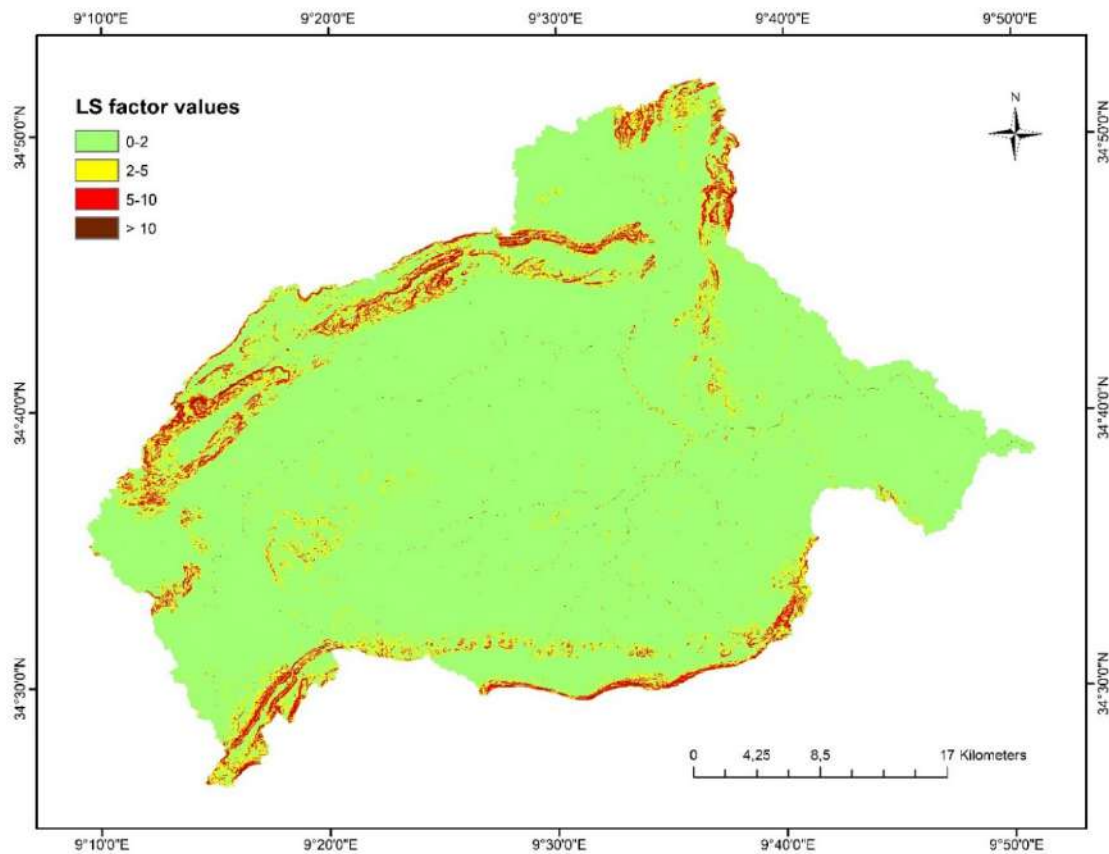


Figure 7 LS factor

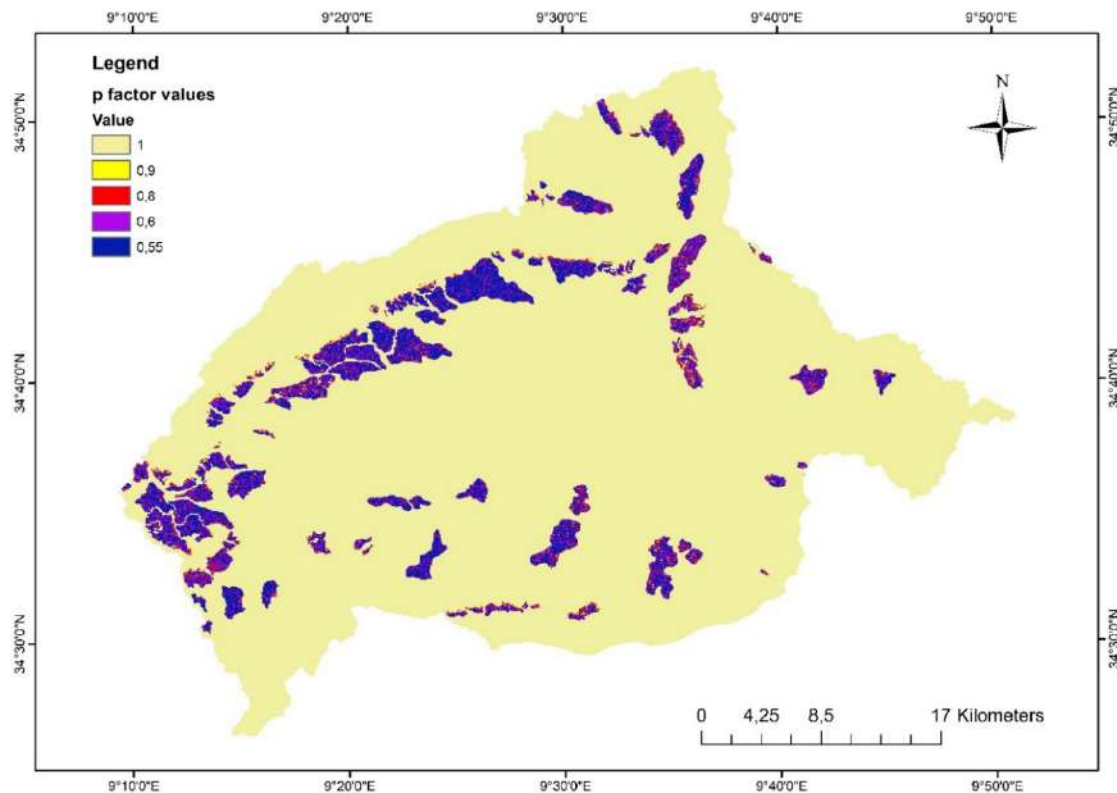


Figure 8 *P* factor

– The slope length and steepness factor *LS*: The watershed values vary between 0 and 836 and they are closely linked to the slope map values, i.e. the mountain areas are those with the highest *LS* values, whereas the central plains have the lowest (Fig. 7). The spatial distribution of the *LS* factor shows that most of the catchment (87%) has values below 2, followed by the moderate slopes, low-reliefs and plateau (7%) which have values between 2 and 5 (Table 7). Then we find the areas of steep slopes of the mountainous border (3.79%) which have values between 5 and 10. Mountain cornices and main rivers (0.41%) have values greater than 10.

Table 7 Classified *LS* factor values

LS factor values classes	Area in ha	%
0–2	129405.10	87.98
2–5	11495.49	7.82
5–10	5570.16	3.79
>10	607.65	0.41

– RUSLE annual total erosion

Seen as a whole, annual soil loss values are evolving between 0 and 731 $\text{tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$ with an average erosion of 3.84 and a standard deviation of 5.08 $\text{tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$.

Erosion map in stretched values based on standard deviation clearly shows a location of elevated values on the mountains of the north, north-west and secondarily of the south of the watershed (Fig. 9). This is logical since the regions mentioned include important mountain systems with narrow inter-mountain depressions and often steep slopes.

Representation of erosion values under the standard RUSLE classification shows the 5–15 $\text{tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$ class as the most characteristic of mountainous regions (4.93%) while the 0 to 1 class is the most widespread in the watershed terrain (78%) followed by the 1 to 5 $\text{tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$ class (15.53%) (Fig. 10) (Table 8).

But when we use the Masson (1971) classification for the semi-arid regions of Tunisia we can see more details in erosion spatial distribution (Fig. 11) (Table 9). Therefore, lands experiencing erosion between 0 and 2 $\text{tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$ represent the

major part of the basin (86.34%). Those losing between 2 and 5 $\text{ton} \times \text{ha}^{-1} \times \text{year}^{-1}$ represent 8.02%, followed by lands suffering soil loss between 5 and 10 $\text{ton} \times \text{ha}^{-1} \times \text{year}^{-1}$ (5.41%). The most critical erosion (above 10) concerns only 1.92%. If we combine the two central classes, we obtain 13.43 (19448.48 ha) of the lands affected by erosion between 2 and 10 $\text{ton} \times \text{ha}^{-1} \times \text{year}^{-1}$.

Table 8 Soil annual loss under standard classes

Erosion classes ($\text{ton} \times \text{ha}^{-1} \times \text{year}^{-1}$)	ha	%
0–1	114139.27	78.84
1–5	22477.60	15.53
5–15	7143.24	4.93
15–50	1009.31	0.70
50–200	10.54	0.01
> 200	1.27	0.0008

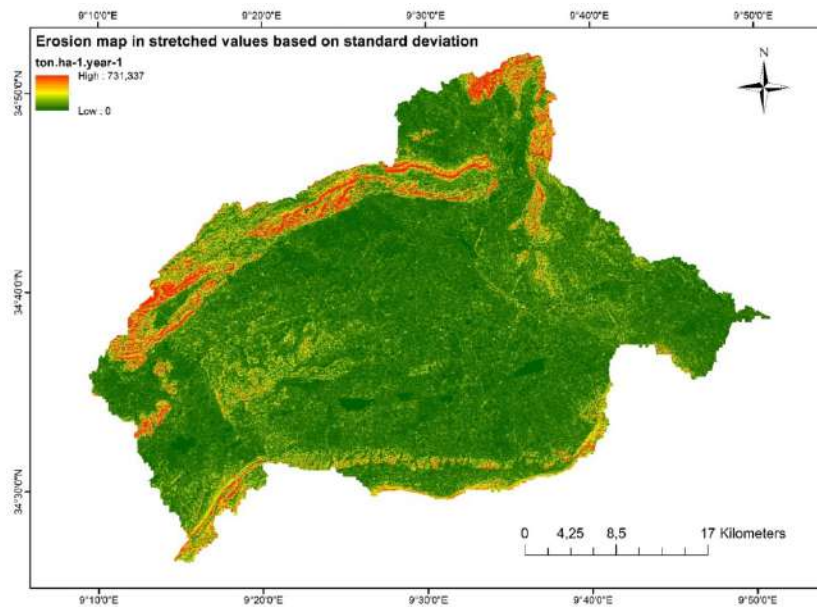


Figure 9 Erosion map in stretched values based on standard deviation

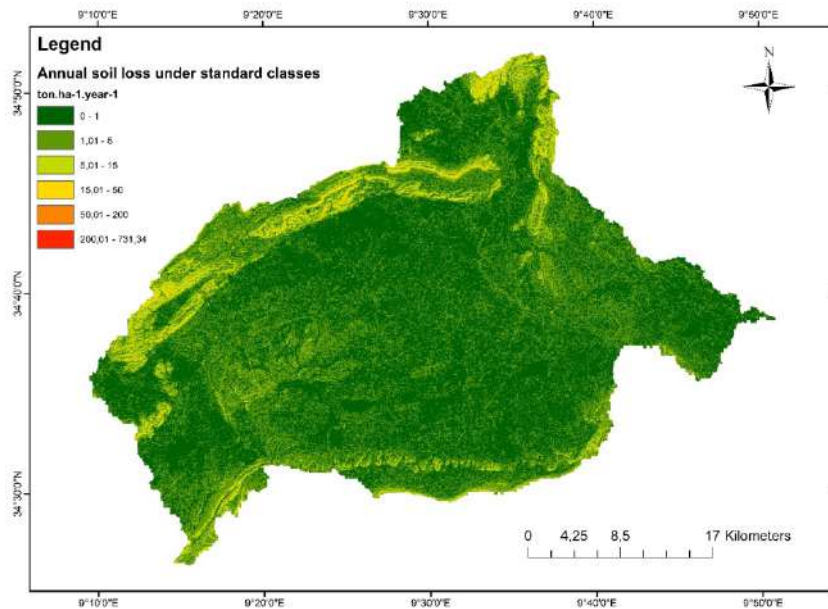


Figure 10 Annual soil loss under standard classes

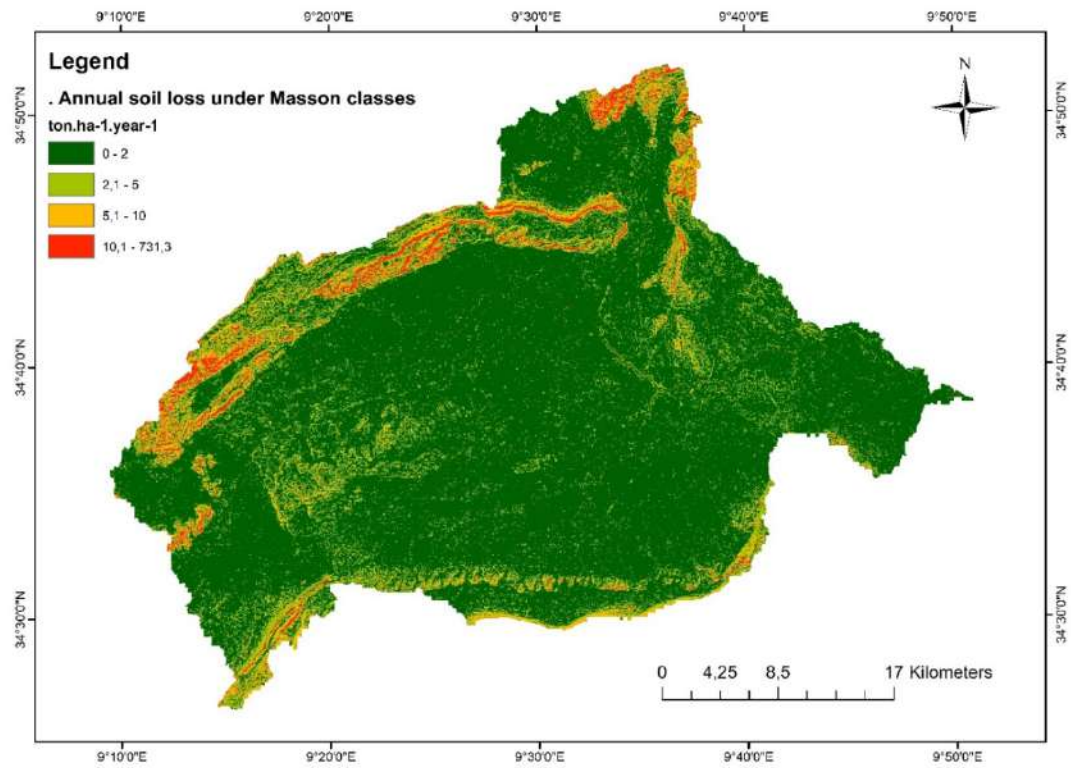


Figure 11 Annual soil loss under Masson (1971) classes for Tunisian semi-arid lands

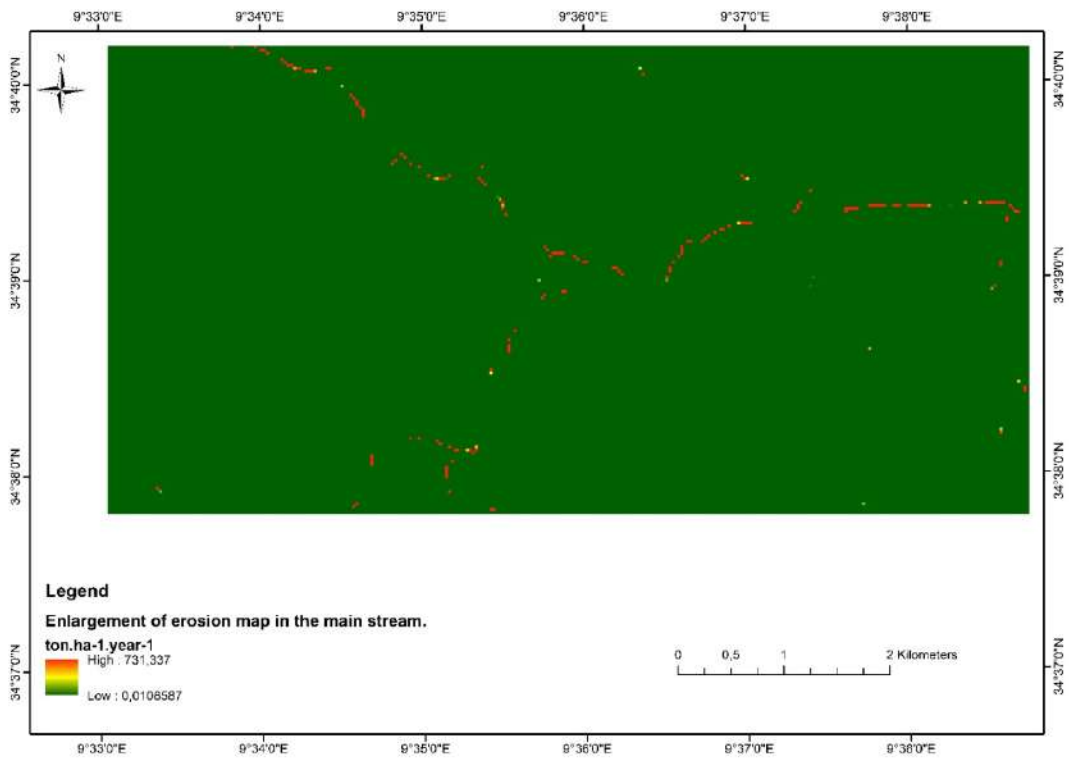


Figure 2 Enlargement of erosion map in the main stream

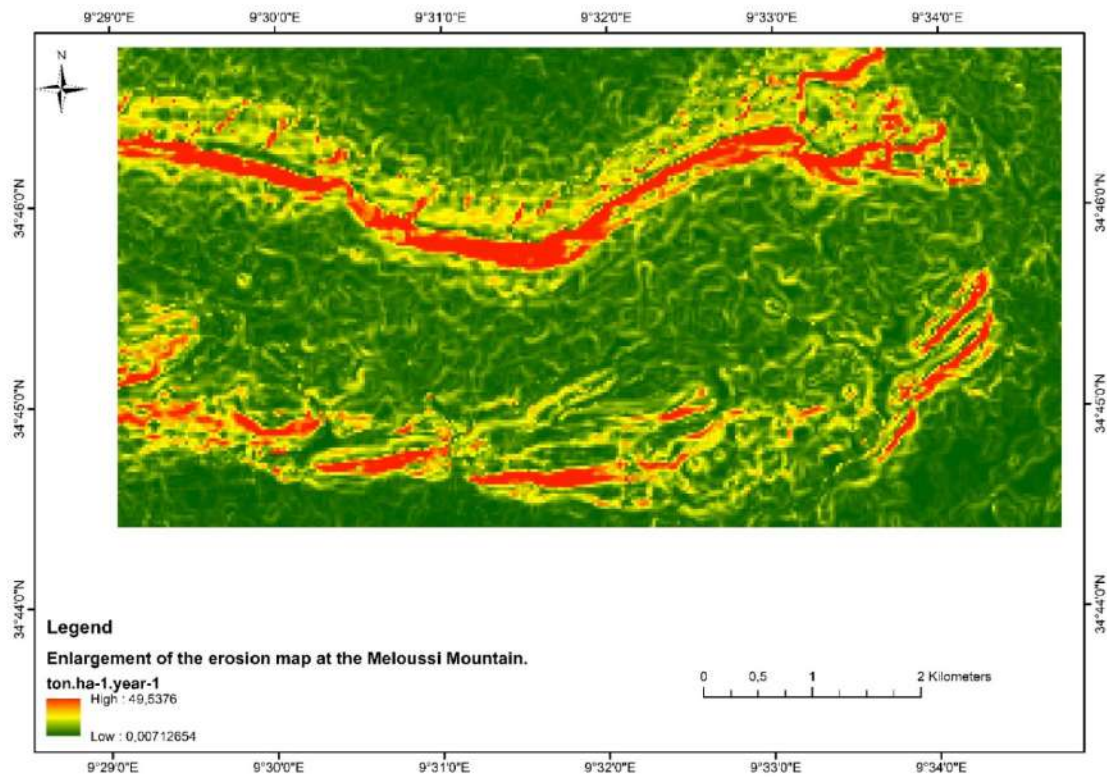


Figure 3 Enlargement of the erosion map at the Meloussi Mountain

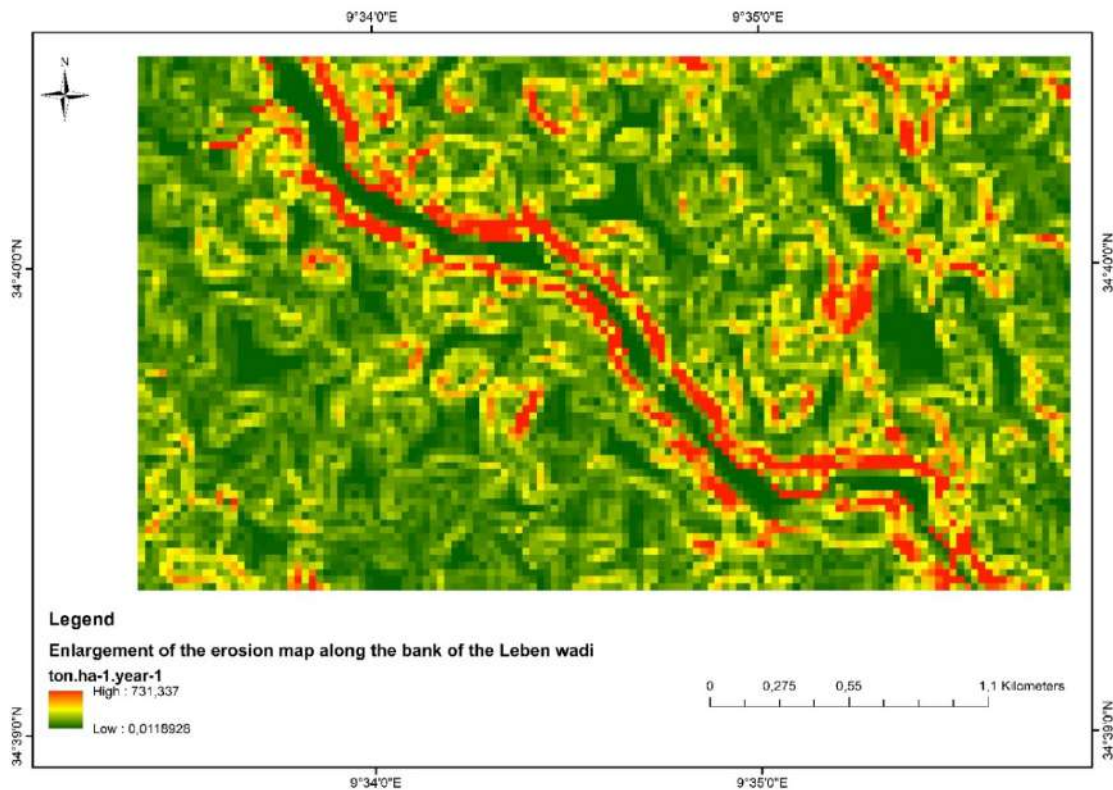


Figure 4 Enlargement of the erosion map along the Leben banks

Table 9 Annual soil loss under Masson (1971) classes for Tunisian semi-arid lands

Erosion classes ($\text{tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$)	ha	%
0–2	124998.74	86.34
2–5	11618.13	8.02
5–10	7830.35	5.41
> 10	2780.37	1.92

The relatively high erosion values recorded on the relief of the catchment borders are clearly related to the presence of soil types with very high erodibility (K above 0.052) which are raw mineral soils, low-developed soils and soils with high erodibility (rendzine). They are also explained by the often high LS values recorded there. On the other hand, reliefs of the catchment margins are rarely cultivated or covered by a vegetal cover. On the contrary, they are experiencing significant overgrazing due to a livestock that is inappropriate with attempts to reforest and improve grazing areas remaining below requirements. Protection against erosion is almost absent in the mountains surrounding the catchment. Reforestation does not seem to succeed and slopes are often denuded denouncing multiple failures: inefficiency of reforestation and protection efforts but also multiple and repeated aggressions by abusive overgrazing.

Comparison with previous studies on erosion quantification may be interesting if authors had used similar formulas and same unit system. Therefore Zante et al., (2003) obtained in the Abdessadok watershed (Tunisian dorsal, Siliana region, 345 mm) an average soil loss of $4.12 \text{ tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$ with a minimum value of 1.75 (the driest year) and a maximum value of 6.03 (the wettest year). These values seem to be concordant with the values recorded here given the similarities between the two catchments.

Moreover, Albergel et al. (2003) reported, at the end of the Hydromed project (study of water erosion in selected basins in the Mediterranean area) and during the annual observation of soil losses, values in the order of $1.8 \text{ tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$ in the Gouazine catchment (Tunisian dorsal, jebel Bargou, 400 mm), $12.2 \text{ tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$ in the Mrichet watershed (Tunisian dorsal, jebel Bargou, 450 mm), $17.5 \text{ tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$ in the Es Senega basin (central Tunisia, Sbiba region, 250 mm), $24.2 \text{ tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$ in the Fidh Ali catchment (central Tunisia, Kairouan region, 300 mm) and $32 \text{ tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$ in the Kamech basin in northeast Tunisia (Table 10).

Later, Ben Chaika et al. (2008) recorded in the Jannet watershed (Tunisian dorsal, Siliana region, 400 mm) a value of $14.8 \text{ tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$ for an average year, with values in the order of $6.5 \text{ tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$ for a dry year and a value of $20 \text{ tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$ for a very wet year.

Table 10 Soil loss in some catchments in Tunisia after Alberger et al., (2003)

Watersheds	Location	Catchment area (ha)	Year of survey	Soil loss ($\text{tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$)
El Gouazine	Central Tunisia, valley in alluvium Pleistocene crusts. (Isohyet 400 mm)	1810	1990	1.8
Es Sénega	Central Tunisia, Cretaceous Mountains. Limestone and gypsum marls (Isohyet 250 mm)	363	1991	17.5
Fidh Ali	Central Tunisia, Eocene anticline. Marls gypsum and lumachelles (Isohyet 400 mm)	238	1991	24.2
Kamech	Central Tunisia, Cap Bon Anticline. Limestone, sandstone and marl (Isohyet 650 mm)	245.5	1993	32.0
M'Richet	Central Tunisia, Bargou anticline. Marls and Cretaceous limestone (Isohyet 450 mm)	158	1991	12.2

Andresson (2010) obtained values of 11.4 and $24.5 \text{ tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$ respectively in the Mrichet and Sabine basins. Gaubi et al., (2017) recorded in the Lebna catchment (North-East of Tunisia) a value

of $24 \text{ tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$. The disparities in erosion values mentioned here are mainly attributable to geographical variations in the R factor. Indeed, rainfall in Tunisia is increasing towards the north-west.

The study of the final erosion map shows that the main streams are not only considered to be areas of high erosion but also have the highest rates of erosion in the basin (Fig. 12). Indeed, if the mountainous sectors record values between 0 and 49 $\text{tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$, (Fig. 13) main streams have values ranging from 0–731 $\text{tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$, the highest in the catchment as shown on the enlargement maps legend. This can be explained by the role of *LS* factor based on slope accumulation and particularly flow accumulation from upstream to downstream of the catchment area. This means that at the level of the main streams and especially their downstream segments, the flows accumulation is maximum, resulting in a maximum of erosion. Thus, the model revision and *LS* factor formulas multiplication have resulted in a consequent improvement in of erosion assessment. Similarly, the enlargement shows that in watercourses the pixels with the most erosion values most often correspond to the banks more than the bed (Fig. 14). This is related to the abrupt change in elevation and slope recorded by the model interpreting it as a flow and material movement.

5. Conclusion

In this study, the objective was to use the RUSLE prediction model to quantify erosion in the Leben catchment. While *R* factor values are significantly low (between 20.58 and 32.59 $\text{MJ} \times \text{mm} \times \text{ha}^{-1} \times \text{hr}^{-1} \times \text{year}^{-1}$), the *K* factor values are substantially high or very high (between 0.04 and 0.075 $\text{tonnes} \times \text{ha} \times \text{h} / \text{MJ} \times \text{mm} \times \text{ha}$). Similarly, the watershed is dominated by bare soils (48%) or olive groves (35%), which do not provide much soil coverage; soil conservation practices have only concerned the piedmont sectors in the catchment. The *LS* factor, closely related to slope maps, reveals mountainous areas as more susceptible to erosion.

Superposition of the different model factors allowed us to compute erosion values and revealed erosion rates ranging between 0 and 731 $\text{tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$ with an average erosion of 3.84 $\text{tonnes} \times \text{ha}^{-1} \times \text{year}^{-1}$. The relatively elevated rates of erosion are mainly found in the relief surrounding the catchment because of the high values of slopes, the skeletal soils which often compose them, but also be-

cause of absence of protection, reforestation or crop management. This means that the process of relief erosion that began in the Tertiary Era continues, taking advantage of soil and vegetation degradation and overgrazing. It means also that, if the mountain slopes are not protected and reforested, erosion will not decrease in the watershed while protection of the piedmont remains relative.

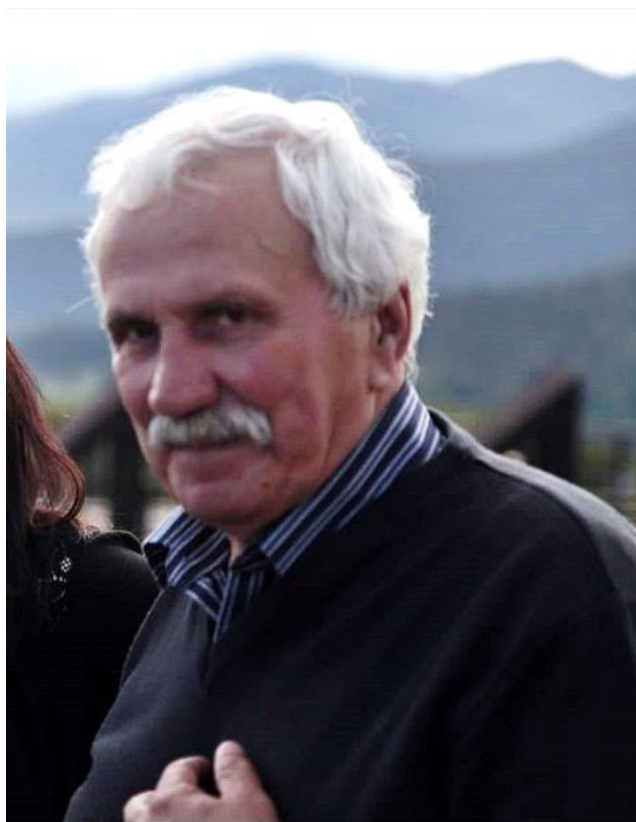
Comparison with previous studies carried out in Tunisia shows similarities in the results with some of them considering the geographical variation of the *R* factor. Analysis of the results at the pixel level showed that the streams and especially their banks record the highest erosion values and this was related to the flow accumulation in the downstream sections.

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Omagiu unui prieten / Tribute to a friend



**Virgil Surdeanu
(1946 – 2018)**

Sâmbătă, 7 iulie 2018. O veste cumplită s-a răspândit în lumea geografică. Decesul neașteptat al bunului nostru coleg și prieten, Virgil Surdeanu. Profesor la Universitatea Babeș-Bolyai, geomorfolog de formație clasică, președinte al Asociației Geomorfologilor din România (2009–2011). S-a stins pe neașteptate, avea doar 72 de ani (n. 15 februarie 1946). Am fost devastată și copleșită de neguri mul-

Saturday, July 7th 2018. Terrible news spread throughout the geographical community. The unexpected demise of our dear colleague and friend, Virgil Surdeanu. Professor at Babeș-Bolyai University, with a classical approach to geomorphology, president of the Romanian Association of Geomorphologists (2009–2011). He died suddenly at just 72 (b. February 15, 1946).

te zile în șir. M-am îndemnat frecvent la plimbări lungi, singuratice, pe cărări liniștite și spații deschise ca să-mi alung fricile neguroase, să-mi adun gândurile, să scormonesc în anii noștri de început, să înțeleg cum a fost curgerea timpului ce ne-a fost dat... Plecarea lui Virgil am simțit-o puternic, ca și cum eu însămi am făcut deja un pas. De ce? Pentru că optsprezece ani (între 1973 și 1991) am trăit împreună devenirea profesiei noastre și consolidarea ei într-o atmosferă atât de spumos descrisă de Profesorul Grasu în "*Strălucirea și crepusculul unei instituții*" (Revista de Geomorfologie, 2017). M-am gândit că cel mai frumos omagiu pe care eu pot să-l aduc lui Virgil este să vorbesc despre noi, cum ne-am șlefuit personalitățile, ciocnindu-ne unii de alții în toți acești ani, astfel încât să semnăm împreună sute de articole științifice, zeci de capitole de carte și chiar cărți întregi, să fim prezenți la cele mai însemnate manifestări științifice naționale (simpozionul anual de geomorfologie fiind un eveniment mult așteptat). De cele internaționale nu era cazul în acele vremi.

Într-o zi de septembrie 1973 a fost prima întâlnire cu Virgil (sau mai degrabă dl cercetător științific Surdeanu). În acea zi m-am pregătit sufletește să iau contact cu viitorul meu loc de muncă, Stațiunea de Cercetări "Stejarul" de la Pângărați. De la gara din localitate am mers mult pe jos pe ulițele satului, apoi am urcat un deal spre niște ziduri groase, maiestose și roase de vreme – o veche ctitorie a lui Alexandru Lăpușneanu.

În curte, pustiu....doar un bărbat înalt, cu o pelerină de sac, cu o pușcă pe braț și un câine flocos la picior (cuminte Pepsi! aud comanda). Aștepta pe cineva. Mai târziu am aflat că acest personaj era celebrul ornitolog Dan Munteanu, viitor academician și directorul Stațiunii de atunci. De el se apropie un fel de om de curte, cu părul vâlvoi, cu niște cizme glodoase, cu un ciocan într-o mână. Nu mă bagă nimeni în seamă. Ignor bărbatul cu cizme (probabil un muncitor cu ziua pe acolo....mai târziu am aflat că era dr. C. Grasu, geolog, viitor profesor la Universitatea A.I. Cuza și mare mânuitor de vorbe de duh ce usturau cumplit când te atingeau!) și mă apropii timid de domnul înalt...Cum pot să ajung la biroul de geomorfologie? (!).Traversez un portic, urc niște scări, străbat un lung coridor și pe o ușă citesc cu

I was devastated and overwhelmed by heavy thoughts for many days. I often took long, lonely walks along quiet trails or in open spaces to drive away the dark fears. I tried to gather my thoughts and rake through our early years, to understand the passage of the time that has been granted to us... It felt as if Virgil's departure was as much a step that I had taken myself. Why? Because for eighteen years (from 1973 to 1991) we lived together, maturing as professionals and building our careers in an environment depicted in such sparkling terms by Professor Grasu in "Strălucirea și crepusculul unei instituții (The shine and twilight of an institution)" (Revista de Geomorfologie, 2017). It dawned on me that the best tribute to Virgil would be to recount about us, how we polished our personalities by clashing with each other throughout all these years and how we ended up working together on hundreds of scientific papers, dozens of book chapters or entire books and participating in the most significant national scientific meetings (among which the annual geomorphology symposium was the most anticipated), as international events were out of our reach in those times.

I first met Virgil (or rather scientific researcher Surdeanu) on a September day in 1973. That day I was preparing myself to become acquainted with my new workplace, Stațiunea de Cercetări "Stejarul" de la Pângărați ("Stejarul" Research Station at Pângărați). From the local rail station I walked for a while on the village streets and then climbed a hill until I reached an old majestic building with thick walls raised by Alexandru Lăpușneanu.

The courtyard was deserted... with the exception of a tall man wearing a coarse-textured cape, a rifle on his arm and a fluffy dog by his foot (commanding it: Good, Pepsi!). He was expecting someone. Later I found out that this character was famous ornithologist Dan Munteanu, future Academy member and the director of the Station in that age. He was then approached by some sort of man of all work with disheveled hair, wearing muddy boots and a hammer in one hand. Nobody noticed me. I ignored the boots-wearing man (probably a hired handyman... whom I would later find out to be dr. C. Grasu, geologist and future professor at Al. I. Cuza University, as well as an unparalleled wielder of witty



Sațiunea de cercetări "Stejarul" Pângărați (1957-1983), zisă "La Castel" în limbajul localnicilor, găzduită într-o veche ctitorie a lui Alexandru Lăpușneanu (1560). În ultimul plan, Dealul Păru, în bază, valea pârâului Pângărați. Drumul de acces serpuiește din vale, urcă pe conul de dejecție al torentului Călugăru, taie perpendicular versantul până la poarta de intrare (foto luată înainte de 1990)

"Stejarul" Research Station at Pângărați (1957-1983), also known as "La Castel" ("At the Castle") by local people, hosted by an ancient building erected by Alexandru Lăpușneanu (1560). In the background, Păru Hill, at the base, Pângărați stream valley. The access road winds from the valley, climbs on the alluvial fan of Călugăru torrent and cuts the slope up to the entrance to the Station (photo taken before 1990)

mare ușurare – Laboratorul de Geomorfologie. Am bătut delicat la ușă. Nimic. Apăs clanța. Deschis. Ah, un hol mic din care se deschid alte două uși. Una cu bare de fier. BDS (mai târziu am aflat că asta înseamnă Biroul de Documente Secrete(!)). Pe cealaltă silabisesc: Dr. Ioan Bojoi, CS Ioniță Ichim, CS Virgil Surdeanu. O voce răsunătoare m-a poftit să intru. De la un birou masiv se ridică un bărbat zâmbitor, cu o mustață pronunțată și un păr negru, stufos, ce-i încadra chipul jovial. Purta un halat alb cu o pipă neagră la buzunarul de la piept. Acum când mă gândesc la el, era doar un tinerel de 27 de ani. Dar atunci, felul extrem de deschis și zgomotos de a mă primi m-a intimidat pe loc. Abia puteam rosti un bună ziua și să mai bâigui ceva, că acest domn cu gesturi largi și cuvinte numeroase m-a așezat pe un scaun și l-a chemat din cealaltă încăpere pe alt personaj care avea să-mi schimbe viața, Ioniță Ichim. "Nene, cică e nevasta lu' alde Rădoane" (N. Rădoane deja se instalase în post). Mic de statură și foarte slab, Ioniță Ichim ne-a dominat

words which stung terribly when directed at you!) and approached the tall man... How can I get to the geomorphology office? (!) I had to go through a portico, climb some stairs and walk through a long corridor to finally read on a door, with great relief – Geomorphology Lab. I knocked gently on the door. Nothing. When I pressed the handle door I noticed it was open, and I entered a small hall with two more doors. One of them was protected by iron bars: BDS (I later found out that meant the Secret Documents Bureau(!)). The other one read: Dr. Ioan Bojoi, CS Ioniță Ichim, CS Virgil Surdeanu. A resounding voice urged me to come in. A smiling man with a heavy moustache and dark, huffy hair and jovial face stood up from a massive desk. He was wearing a white lab coat with a black pipe in his chest pocket. Now that I picture him, he was just a 27-year old. But at the time, his very open and noisy welcome was quite intimidating to me. I could barely greet him and stutter a few words, when this man with ample gestures and numerous words sat

pe toți în anii lungi ce-au urmat. Nimeni dintre noi nu-i putea spune pe numele mic. Cel mai inventiv în găsirea de expresii și titluri de adresare era Virgil („șefu’, nene, maestre, boss, nea...” ș.a).

Din acea zi am intrat în malaxorul modelării superioare a spiritului nostru. Veniți de pe băncile școlii cu câteva deprinderi și niscaiva cunoștințe generale, plus câteva talente native, am aflat curând ce uriașă ne era ignoranța. Virgil era neîntrecut la calcule matematice mentale. Pe teren, în special, își manifesta cu celeritate talentul, calculând distanțe, suprafețe, volume, imaginând modele complicate de dislocare de roci și rate de eroziune. La început, nu l-am prea crezut. Se cheltuia timp pentru verificarea



Virgil Surdeanu și Nicoale Radoane (în anul de grație 1974) îl încadrează pe proaspătul doctor în geografie, Do Hung Than, din Vietnam

Virgil Surdeanu and Nicoale Radoane (in the year of grace 1974), framing the new PhD in Geography, Do Hung Than, from Vietnam

me down on a chair and called another man from the other room, who would go on to change my life, Ioniță Ichim. "Nene, this is the Rădoane's wife" (N. Rădoane had already settled in the post). Rather short and very slender, Ioniță Ichim dominated us all in the long years that followed. None of us even thought to thee and thou him. The most resourceful in finding expressions and ways of addressing him was Virgil: „șefu’, nene, maestre, boss, nea...” a.s.o.

From that day began a new era in our lives when our spirits where shaped. Fresh university graduates, with slim skills and some general knowledge plus certain native talents, we would soon find out the extent of our ignorance. Virgil was unsurpassed when it came to mental mathematic calculations. In the field, in particular, his talent was often put to good use by calculating distances, areas, volumes, imagining complicated patterns of rock displacement and erosion rates. At first we were skeptical and spent enough time checking his computations and measuring distances using the compass, the roulette or the double-step. However, we soon gave up as we noticed how small the differences were. His rational thinking helped him to fully understand a geomorphological process as unpredictable as landslides. He dreamt of elaborating a mathematical prediction model for landslide triggering and evolution. In order to achieve that he also relied on engineering approaches which focused on finding direct and immediate solutions for slope stabilization. I was fascinated by the long passionate discussions Virgil had with the best practitioners in slope dynamics in the evenings, on the gorgeous terrace of the Station after a fruitful day in the field. These included R.J Bally, N. Borșariu, P. Stănescu, who had been persuaded by Virgil to visit Pângărați and observe the perimeters with landslides from Bistrita valley. The results were soon seen in the array of articles published and his focus on discerning the internal mechanisms of the landslide body is best illustrated by the book "Geografia terenurilor degradate. Alunecările de teren" (Geography of degraded lands. Landslides, 1998).

The long periods we spent together working in the office, on field campaigns or at family events meant that we had gone through many discussions, loud exchanges (conversations were loud in our

calculelor lui, măsurând distanțele cu compasul, ruleta sau la dublu-pas. Ne-am lăsat curând păgubași când vedeam ce mici diferențe erau. Gândirea lui rațională l-a ajutat mult în cunoașterea unui proces geomorfologic atât de imprevizibil cum sunt alunecările de teren. Visa mult să poată realiza un model matematic de predicție a declanșării și evoluției alunecărilor de teren. Pentru asta a creat punți de legătură cu abordările ingineresti asupra acestor procese, cei care erau chemați să vină cu soluții directe și imediate în consolidarea versanților. Eram fascinată de discuțiile lungi cu patimă pe care Virgil le avea cu cei mai buni practicieni în dinamica versanților, seara, pe superba terasă a Stațiunii, după o zi de teren fructuoasă. Aceștia erau R.J Bally, N. Borșariu, P. Stănescu, convinși de Virgil să vină la

geomorphology office), sadness and misfortunes, lack of money, huge mugs of "nechezol" (a dubious mixture of chicory, barley and coffee powder), all seasoned with endless childhood stories, tales of the astral moments of the teen years, funny stories as well as most dramatic... we shared them all with such openness and sincerity which would seem so unusual today.

The mandatory start of a typical work day would be a cup of coffee or tea, while conversation topics were most often initiated by Ioniță Ichim. After we moved to Piatra Neamț (1983), Virgil changed his residence to Bacău and used to commute by train. It was annoying and even horrendous during winters.



O campanie de teren dura între 7 și 10 zile, de cele mai multe ori ne deplasam cu trenul, locuiam la săteni, cămine muncitorești, iar uneltele de lucru (în cazul de față șubler, cântar, ciururi, lopeți, cazma, nivelă, stadie, rigle, sfoară, pungi pentru probe) erau ambalate în hârtie groasă, legate bine cu sfoară și repartizate fiecăruia dintre noi. Aici suntem pe malul r. Buzău la Varlaam. De la dreapta la stânga, Virgil Surdeanu, Ioniță Ichim, Maria Rădoane (eu mereu cu caietul în mână, mi se mai zicea "duduia de la cancelarie"), iar la aparatul foto, N Rădoane, Anul 1984

A field campaign lasted between 7 and 10 days. Most of the times we travelled by train, stayed at villagers' houses or workers' dormitories, and our tools (including calipers, weighing scales, sieves, shovels, pickaxes, level, stadia rod, rope, sample bags) were all packed in thick paper bundled together with rope and assigned to each of us. Here we were on the banks of river Buzău at Varlaam. Right to left: Virgil Surdeanu, Ioniță Ichim, Maria Rădoane (me, always carrying a notebook in my hand, also known as the "lady from the chancellery") and N Rădoane handling the camera, 1984

Pângărați și să vadă perimetrele de alunecări de pe valea Bistriței. Rezultatele s-au văzut curând în suita de articole publicate, iar accentul pe deslușirea mecanismelor interne ale unui corp de alunecare este bine ilustrat în cartea lui *"Geografia terenurilor degradate. Alunecările de teren"* (1998).

Timpii lungi petrecuți împreună, la biroul de lucru, în campaniile de teren, la evenimente de familie au însemnat multe discuții, schimburi zgomotoase de cuvinte (se vorbea tare în cabinetul nostru de geomorfologie), supărări și neîmpliniri, lipsă de bani, câni imense de "nechezol" (un amestec dubios de cicoare, orz și un praf de cafea), totul stropit cu povești nesfârșite din copilăria fiecăruia, momentele astrale ale adolescenței, cele mai hazlii întâmplări, dar și cele mai dramatice... toate ni le împărtășeam deschis și cu o sinceritate de care astăzi parcă m-aș jena.

O zi de lucru începea neapărat cu o cană de așa zisa cafea sau de ceai, iar subiectele de discuții erau inițiate cel mai adesea de Ioniță Ichim. După ce ne-am mutat în Piatra Neamț (1983), Virgil își schimbă domiciliul în Bacău și făcea o navetă enervantă cu trenul. Iarna, cel puțin, era înfiorător. "Vă rog frumos, jumătate de oră să mă lăsați în pace, nu mă întrebați nimic, nu mă rugați să vorbesc că nu o să vă placă ce auziți. Sunt înghețat ca un c...t". Încet, încet, atmosfera se anima în jurul subiectului/lor în discuție. Era vorba de o lectură din seara precedentă a unei cărți (Nichita Stănescu era idolul nostru), a unui articol, a situației din Orientul Mijlociu, ori vreo știre de la Europa Liberă (mai ales după momentul Mangalia 1983¹), dar cel mai adesea dezbăteam geomorfologie.

Folosind fluturași special tipăriți (cât o carte poștală) pe o hârtie de proastă calitate, trimiteam în lumea largă cereri de extrase ale articolelor științifice, cu adrese aproximative ale autorilor. Se întorceau cu succes cam 5%. Singurul dintre noi care ieșea în afară în fiecare an era Ioniță Ichim, iar la întoarcere era burdușit de comori geomorfologice. Așa am avut șansa să dezbătem împreună cele mai incitante idei și concepte din geomorfologie, aproa-

"I am asking you kindly, please leave me alone for half an hour, don't ask me anything, don't expect me to talk because you will not like what you hear. I'm frozen like s..t". Slowly, the atmosphere would become animated around the topic/s being discussed, whether it was a reading from the previous evening (Nichita Stănescu was our idol), an article, the situation in the Middle East, some news on Free Europe Radio (particularly after Mangalia in 1983²), or, most frequently, geomorphological debates.

We would send out in the world requests for scientific article extracts using postcard-sized flyers specially printed on low-quality paper with approximate addresses of the authors. The successful return rate amounted to about 5%. The only one of us who ever traveled abroad each year was Ioniță Ichim, who, upon return, brought back a plethora of geomorphological treasures. This was how we had the opportunity to discuss together the newest, most exciting ideas and concepts in geomorphology almost as soon as they were being introduced in publications (denudational balance, A Jahn; time, space and causality in geomorphology, SA Schumm; dynamic basis of geomorphology, AN Strahler; dynamic equilibrium theory, JT Hack; the systems approach in physical geography, R Chorley and B Kennedy etc). Oftentimes awkwardly, with youthful bravery or crazy nerve, we began to interpret our results in the spirit of these ideas, whose influence can be found in our publications from that era and our doctoral dissertations. Now, when I summarize the 18 years of ups and downs, I am amazed how we were at no point divided by envy, hate or revenge, or by the well-known collegial jealousy so copiously present in academic environments. But by no means were we modest angels, as we often argued with reproaches and cries (particularly on my side), with crude and blatant southern expressions (mostly on Virgil's side), or subtle whips with philosophical references (mainly from our mentors, I Ichim and C Grasu, the "Lowland" and the "Mountain", to quote the latter). Nevertheless, I believe there was plenty of respect between us, well-ce-

¹ Este vorba de discursul lui Nicolae Ceaușescu din 3 august 1983 care trasa noile directive "politico-educative" în cultura românească

² This is a reference to the speech made by Nicolae Ceaușescu on August 3, 1983, thus establishing the new „political-educational” directives in the Romanian culture

pe în același timp cu lansarea lor în publicații (bilanțul denudațional, A Jahn; timp, spațiu, cauzalitate în geomorfologie, SA Schumm; baza dinamică a geomorfologiei, AN Strahler; teoria echilibrului dinamic, JT Hack; abordarea sistemică în geografia fizică, R Chorley și B Kennedy etc). Cu stângăcie, cu tupeul tinereții, cu un curaj nebun am început să interpretăm rezultatele noastre în spiritul acestor idei (influența se regăsește în publicațiile noastre din acea vreme și în tezele de doctorat).

Acum când comprîm cei 18 ani de suișuri și coborâșuri, mă întreb cum de nu ne-a încercat nici o clipă patima invidiei, a urii și răzbunării, binecunoscuta pizmă intercolegială care se distribuie generos în mediile academice. Nici noi nu eram niște îngerași pudibonzi, ne certam gospodărește, cu reproșuri și plânsete (mai ales din partea mea), expresii grosiere și vorbe neaoșe sudiste (mai ales din partea lui Virgil), biciuiri subtile cu inserții filozofice și citate docte adaptate situațiilor (mai ales din partea măștrilor noștri, I Ichim și C Grasu, "Șesul" și "Muntele", ca să-l citez pe ultimul). Dar cred că era un respect între noi, consolidat în timp prin mâna întinsă la nevoie și prin generozitate. Virgil era, de departe, cel mai bine situat prin originile sale. Aveam grijă mereu să-l periem, amical, că e fiu de preot. Însă și poveștile copilăriei lui, petrecute în arșița Bărăganului, nu se deosebeau de ale noastre, copii de țărani, de la deal sau de la munte.

Pe un om îl judeci cel mai bine atunci când îl vezi în situații extreme. Era în 1975 pe valea râului Bârlad. Se ridicau topometric mai multe sectoare de albie minoră. Debitele râului erau mari (după o serie de viituri), astfel încât barca de cauciuc nu ne ajuta mereu. Cineva trebuia să stea în apă cu stadia. Nimeni nu știa să înoate. Singurul, Virgil. Proaspăt căsătorit, cu o verighetă mândră pe deget, ne strigă să ne închinăm pentru el să nu-l ia curentul. Apa, extrem de murdară. În amonte, undeva, era o topitorie de în. Pe apă pluteau mici plauri cu deșeuri vegetale. Virgil reușește să dea puncte cu stadia pe întreg sectorul de râu, în uralele noastre de încurajare. La un moment dat un plaur se lipește de el. Virgil încearcă să-l îndepărteze. Un urlet monstruos se aude. "E un porc moooort!!!!" Lasă totul și iese la mal. Dezgustul și greața ne-a cuprins instantaneu.

mented in time by generosity and the help granted when needed. Virgil was by far the best endowed due to his origins. We always lauded him amicably for being the son of a priest. However, the stories of his childhood spent in the heat of Bărăgan were no different from ours, children of peasants from the hills or the mountains.

A man can best be assessed in extreme situations. It was back in 1975, on the valley of river Bârlad, during a topographical survey of several channel reaches. The streamflow was high (ensuing a series of floods), such that the rubber boat was of little help. One of us had to stand in the water holding the stadia rod. The only one who could swim was newly-wed Virgil, proudly wearing his wedding band, shouting at us to pray for him so that he wouldn't be swept by the current. The water was extremely dirty with floating islets of vegetal waste, as somewhere upstream there was a textile fiber processing plant. Virgil managed to survey the entire river reach in our cheers of encouragement. At some point, however, one of the islets stuck to him, while he was trying to remove it. A monstrous howl was then heard: "It's a deaaaad pig!!!". He left everything and came out of the water. We were all instantly filled by disgust and nausea. Virgil was vomiting in between expletives and heavy words. He needed to vent. We washed with whatever water we had at our disposal. Soon, another shout was heard. "I lost the wedding band! What will I tell my wife?", he cried hopping in despair with his water-wrinkled hands stretched forward. Cohorts of mosquitos took greedy bites of us. Someone maintained a fire of rotting wood. Then, slowly, he began to laugh at the situation and we recounted this event for a long time. A year later (1976) we published a great article in Revue Roumaine de Geographie, which was at the time the most recognized journal in our field, presenting the results of that extremely difficult and dangerous, as well as slimy work.

That was the Virgil Surdeanu I knew, a dear, great friend, a generous human being, a man of his word. In 1991 we each went his own way, headed for an Alma Mater. In the important days of the years we found each other on the phone, when I had to answer in great detail to Virgil's questions re-

La Virgil s-a lăsat cu vomă și ocări și vorbe grele. Trebuia să se descarce. Il spălăm cu ce apă aveam. După care ...alt strigăt. "Nu mai am verigheta! Ce-i spun eu nevestei?" și deznădăduit țopăie cu mâinile întinse, zbârcite de apă. Roiuri de țânțari mușcau lacom din noi. Cineva întreținea un foc din putregaiuri. Apoi, încet, încet a făcut haz de necaz și mult timp am povestit această întâmplare. Un an mai târziu (1976) am publicat împreună un frumos articol în *Revue Roumaine de Géographie* (la vremea aceea cea mai puternică revistă din bransă) cu rezultatele acelei munci extrem de dificile și periculoase și chiar ...slinoase.

Acesta a fost Virgil Surdeanu pe care l-am cunoscut eu, un prieten bun și drag, un om generos, un om de cuvânt. În 1991 ne-am despărțit, fiecare îndreptându-se, pe rând, spre câte o Alma Mater. În zilele importante ale anului ne găseam la telefon când trebuia să răspund lui Virgil detaliat ce face fiecare membru al familiei și să transmit la fiecare cele cuvenite. Apoi, înșira o listă lungă de nume să întrebe de fiecare cum o duce și pe unde mai trăiește. Mereu încheia apelul cu "Mi-e dor de Pângărați, ooo, nu știți voi cât de dor îmi e!".

Maria Rădoane
6 septembrie 2018

garding the whereabouts of each family member and to convey his greetings to each of them. Then, he would enumerate a long list of names, wondering how everyone was doing and faring. In the end, he would invariably end his call by saying "I miss Pângărați, oooh, you have no idea how much I miss it!"

Maria Rădoane
6 September 2018

International Conference

From field mapping and landform analysis to multi-risk assessment: challenges, uncertainties and transdisciplinarity Buzău, România, May 16–20, 2018

The International Conference *From field mapping and landform analysis to multi-risk assessment: challenges, uncertainties and transdisciplinarity* took place in Buzău, Romania, between May 16 and 20, 2018. Organized by the Institute of Geography (Romanian Academy) and the Faculty of Geography (Univ. of Bucharest) under the auspices of the International Association of Geomorphologists (IAG) and the Geomorphologists Associations from Romania, Italy, France, Belgium and Greece, the Conference gathered two scientific events with a long history – the 19th Joint Geomorphological Meeting Italy–Romania–France–Belgium–Greece and the 34th Romanian National Symposium on Geomorphology, dedicated to 50 years of activity of Pâtârlagele Natural Hazards Research Center.

The Conference gathered 80 participants from 8 countries (Italy, France, Belgium, Greece, Japan, Canada, Poland and Romania), for two days of keynote lectures, oral and poster presentations alternating with two field trips in the Vrancea seismic region (the Curvature Curvature sector of the SE Carpathians). The conference was meant to encourage participation from all geomorphology fields, as well as inter- and transdisciplinary research, to provide a high-level but in the meantime comfortable and friendly framework, able to promote active discussions and challenging scientific debates. The topic of the conference was outlined by 7 keynote lectures (Emil Sever Georgescu: *Seismic landscape of Romania: challenges of a specific paradigm*; Monique Fort: *Earthquakes impact on geomorphology (landforms and processes) in the Himalayas at different time scales*; Takashi Oguchi: *Digital ele-*

vation models and their applications in geomorphology: historical review and future perspectives; Alessandro Chelli: *The activities of AIGeo Working Groups towards a new applied geomorphic map*; Olivier Dewitte: *Landslide hazard assessment in an urban-sprawling context. A geomorphological approach in Bukavu, DR Congo*; Florin Pendea: *Transdisciplinary approaches to advancing knowledge of glacio-isostatic rebound in Northern Canada*; Konstantinos Vouvalidis: *Hydromorphological assessment of suburban torrents for flood protection: the flood event of West Attica, Greece*). Besides these, 23 oral plenary presentations brought into discussions issues like tectonic jointing/control on large landslides occurrence/evolution, seismic landslides, the use of dendrochronology in the study of shallow and deep-seated slides or debris flows, glaciated karst, permafrost distribution modeling, improved landslide scarp detection, geoarchaeology and anthropogenic geomorphology, flood risk and reservoir siltation, anthropogenic subsidence, paleogeographic transformations, medium-term morphodynamics, natural variability and human influence in the evolution of deltas, variability in sediment budgets at river mouths, zoogeomorphic impact of animals in alpine environments or the use of multi-temporal satellite interferometry in landslide detection and monitoring, found their position in the multi-hazard paradigm. During a 3-hours long session, including 2-minutes madness presentations, 25 posters with the following topics were intensely debated: fluvial multi-temporal evolution and risks (from fluvial terrace formation and channel forming and discharge to financial losses); snow avalanche

susceptibility mapping; paleo-reconstruction of floodplains; coastal evolution patterns and near-shore sandbars positioning; mountain climate variability and its influence on permafrost and frost weathering; geomorphological assessments of major landslides and their role in relief evolution; neuro-fuzzy systems and neural network technology for drainage sub-basins classification and slope erosion by runoff analysis; changing Alpine glacier forelands; hillslope – river channel coupling, etc.

Two field trips have been organized on Friday, May 18 and Sunday, May 20, which aimed at providing a detailed image on the multi-hazard hotspot of Buzău Carpathians and Subcarpathians, on the conceptual framework of single-to-multi-

hazard assessment and on the inter- and transdisciplinarity characteristics of geomorphology within a multi-risk approach.

A multi-hazard hotspot of Europe, Vrancea seismic region is offering the perfect framework for observing, discussing and evaluating a multi-risk environment, where forms and processes, alongside prediction, modeling and validation of single and multi-hazards are fitting perfectly within an extremely complex yet challenging system of approaches. The field trips were focused on high magnitude-low frequency landslides (deep seated, mainly dormant, with large reactivation potential) and their climatic and seismic triggers. Aspects concerning deep-focal seismicity, litho-structural conditioning,



Some of the key case-studies discussed during the field trip: Siriș reservoir, Chirlesti earth flow, Groapa Vantului complex landslide.



Conference snapshots: field explanations at Balta landslide (upper, left), posters presentation (upper, right), Keynote lectures: transdisciplinarity, reviews and perspectives in geomorphology (lower; Assoc. Prof. Florin Pendea, Canada and Prof. Takashi Oguchi, Japan)

active and blind faults have been also discussed. The human impact on the environment was emphasized through case-studies like road and reservoir construction. Morphodynamic mapping, landslide susceptibility and hazard scenarios, numerical and statistical models have been detailed through field examples.

Following the active discussions held during both lectures and field trips, we are positive that the conference opened new research ideas and collaboration initiatives. The 20th edition of the Joint Geomorphological Meeting will take place in Greece, organized by the Hellenic Committee for Geomorphology and Environment.

Mihai Micu, *Institute of Geography, Romanian Academy*

Romanian Young Geomorphologists Day 2018 – 1st Edition Cluj–Napoca, Romania, July 13, 2018

The first edition of *Romanian Young Geomorphologists Day* was held on Friday, 13th of July 2018, at the Faculty of Geography, Babeş–Bolyai University of Cluj–Napoca, under the organization of the Romanian Association of Geomorphologists and the auspices of the International Association of Geomorphologists (IAG). A total number of 25 participants from the Romanian Academy, University of Bucharest, Alexandru Ioan Cuza University of Iaşi, Babeş–Bolyai University and Gheorghe Lazăr High School (Cluj–Napoca) attended this event.

During the official opening, Dr. Mihai Micu (President of the Romanian Association of Geomorphologists, IAG Secretary General) gave an overview on the support provided by the International Association of Geomorphologists for the organization of such events, at both national level as well as at the international one, an International Day of the Young Geomorphologists being among the priorities of the IAG's Training Programme. The opening was followed by two sessions of oral presentations gave by Romanian young geomorphologists, with focus on

modern research methods applied for the analysis of various geomorphological processes such as SAR remote sensing techniques, reconstruction of paleo–river systems, landslide mapping and investigation, glaciation and deglaciation reconstruction, dendrogeomorphic reconstructions. In the week before the event, a Summer School entitled *Dendrochronology applied in natural hazards assessment* was organized by the team of Laboratory of Dendrochronology from the Faculty of Geography, Babeş–Bolyai University. The participants discussed their results as an extended oral presentation during the second session of the Romanian Young Geomorphologists Day. The concluding discussions emphasized the role of such meetings to be held yearly as a necessity for the young geomorphologists career development; the perspectives of organizing the next editions of the Romanian Young Geomorphologists Day during the Romanian National Symposium on Geomorphology was unanimously acknowledged.



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