

**ROMANIAN ASSOCIATION OF GEOMORPHOLOGISTS**

# **REVISTA DE GEOMORFOLOGIE**

**22**



*editura universității din bucurești®*  
2020

# Revista de Geomorfologie 22/2020

## Editor-in-Chief:

Maria Rădoane, University of Suceava, Romania

## Guest Editors:

Olimpiu Pop, Babeş Bolyai University of Cluj-Napoca, Romania

Armelle Decaulne, University of Nantes, France

## Editorial Board:

Achim Beylich, Geological Survey of Norway

Armelle Decaulne, Université de Nantes, France

Aurel Perşoiu, Institute of Speology Emil Racovita, Romania

Ciprian Mărgărint, Al I Cuza University of Iaşi, Romania

Dan Dumitriu, Al I Cuza University of Iaşi, Romania

Daniel Germain, Université du Québec à Montréal, Canada

Dănuţ Petrea, Babeş Bolyai University of Cluj-Napoca, Romania

Floare Grecu, University of Bucharest, Romania

Florin Pendea, Lakehead University, Orillia, Canada

Florin Tătui, University of Bucharest, Romania

Hans-Balder Havenith, Université de Liège, Belgium

Ian Evans, University of Durham, United Kingdom

Ion Ioniţă, Al I Cuza University of Iaşi, Romania

Jean-Philippe Malet, Université de Strasbourg, France

Laura Comănescu, University of Bucharest, Romania

Lucian Drăguţ, West University of Timisoara, Romania

Marta Jurchescu, Institute of Geography, Romania

Mauro Soldati, Università degli Studi di Modena e Reggio Emilia, Italy

Mihai Micu, Institute of Geography, Romania

Mihai Niculiţă, Al I Cuza University of Iaşi, Romania

Mircea Voiculescu, West University of Timisoara, Romania

Nicolae Cruceru, Spiru Haret University, Romania and Institute of Speleology Emil Racovita, Romania

Olimpiu Pop, Babeş Bolyai University of Cluj-Napoca, Romania

Paola Reichenbach, Istituto di Ricerca per la Protezione Idrogeologica, Consiglio Nazionale delle Ricerche, Italia

Petre Urdea, West University of Timisoara, Romania

Piotr Gebica, University of Rzeszów, Poland

Philip Deline, Université de Savoie, France

Sandu Boengiu, University of Craiova, Romania

Sipos Gyuri, University of Szeged, Hungary

Sunil Kumar De, North Eastern Hill University, India

Zofia Raczkowska, Institute of Geography and Spatial Organization, Poland

## Technical Board:

Journal manager: Florin Tătui, University of Bucharest, Romania

Layout editor: Sanda Roşca, Babeş-Bolyai University, Cluj-Napoca, Romania

Proofreading: Florin Zăinescu, Răzvan Popescu, University of Bucharest, Romania

Copyeditor: Mirela Vasile, Research Institute of the University of Bucharest, Romania

## Cover Photo:

### DTP: Meri Pogonariu

Address: Şos. Panduri, 90-92, Bucureşti – 050663, România,

Phone/Fax: (0040) 021.410.23.84, E-mail: editura.unibuc@gmail.com,

Online bookshop: <http://librarie-unibuc.ro>,

Sale point: Bd. Regina Elisabeta, nr. 4-12, Bucuresti,

Tel. (0040) 021.314.35.08/2125,

Web: [www.edit.unibuc.ro](http://www.edit.unibuc.ro)

ISSN 1453-5068

ISSN online 2285-6773

## CONTENTS

### ARTICLES

|                                                                                                                                                                                                                                         |    |
|-----------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------------|----|
| <b>The effects of biocrusts on soil parameters in a semi-arid pediment at north-eastern Iran</b><br>Atoosa GHOLAMHOSSEINIAN, Adel SEPEHR, Iraj EMADODIN .....                                                                           | 5  |
| <b>Analysis of the difference in depths and variation in slope steepness of the Sunda Trench, Indonesia, east Indian Ocean</b><br>Polina LEMENKOVA.....                                                                                 | 21 |
| <b>Investigating land surface deformation using InSAR and GIS techniques in Cluj–Napoca city’s most affected sector by urban sprawl (Romania)</b><br>Anna–Hajnalka KERÉKES, Szilárd Lehel POSZET, Călin BACIU .....                     | 43 |
| <b>Rock glaciers in mixed lithology: a case study from Northern Pirin</b><br>Emil GACHEV .....                                                                                                                                          | 61 |
| <b>Snow–avalanche history reconstructed with tree rings in Parâng Mountains (Southern Carpathians, Romania)</b><br>Corina TODEA, Olimpiu Traian POP, Daniel GERMAIN .....                                                               | 73 |
| <b>In search of marine terraces from Central Dobrogea (Casimcea Plateau, Western Black Sea). Preliminary inquiries on distribution, extension and vertical crustal movements</b><br>Alexandru BERBECARIU, Alfred VESPREMEANU-STROE..... | 87 |



---

# The effects of biocrusts on soil parameters in a semi-arid pediment at north-eastern Iran

Atoosa GHOLAMHOSSEINIAN<sup>1</sup>, Adel SEPEHR<sup>1\*</sup>, Iraj EMADODIN<sup>2</sup>

<sup>1</sup>Department of Desert and Arid Zones Management, Ferdowsi University of Mashhad, Azadi Square, Mashhad, Iran

<sup>2</sup>Institute for Crop Science and Plant Breeding, Christian Albrechts University of Kiel, Hermann Rodewald Str. 9, 24118, Kiel, Germany

Received 13 July 2020; Revised 17 October 2020; Accepted 20 October 2020

\*Correspondence to: Adel SEPEHR, e-mail: [adelsepehr@um.ac.ir](mailto:adelsepehr@um.ac.ir)

---

## ABSTRACT

In arid and semi-arid environments, soils are very fragile. Their degradation rapidly reduces the biological potential and disrupts the ecological and socio-economic balance. Living within a few millimetres of the soil surface, biocrusts are organised into communities of biota and play an important role in soil stability and preservation. They are mainly composed of cyanobacteria, lichens, mosses, fungi, eukaryotic algae and other heterotrophic bacteria. Soil-biocrusts interaction is therefore very important and a good knowledge of this interaction allows a better management of soils, especially in arid and semi-arid environments. Thus, the link between cyanobacteria species and physicochemical parameters and soil mineralogy was studied in two geomorphological zones of northeast of Iran. Cyanobacteria are the main photosynthetic component of biocrusts. Samples were collected in summer along a linear transect by using 50 x 50 cm quadrates for each study zone. Individual mineral soil particles were analysed by scanning electron microscopy (SEM). The amounts of Na, K, Ca and Mg as well as electrical conductivity and soil organic carbon are higher in the presence of biocrusts (Takhte Soltan site) than in their absence (Sabzevar Playa site). Biocrusts increased levels of organic carbon, pH, calcium carbonate, exchangeable sodium and potassium percentages. Results indicated that soil properties regarding soil organic carbon, CaCO<sub>3</sub> and amount of clay changed among biocrust sand bare soils. This research considers further attention to biocrusts because of their multiple ecosystem roles and their importance in the constitution of dryland ecosystems.

**KEYWORDS** arid and semi-arid; biocrusts; cyanobacteria; soil physicochemical properties; mineral erosion; Iran

---

## 1. Introduction

Biological soil crusts (BSCs) are the dominant biological surface features in many arid and semiarid

area supporting vascular plant communities (Maier et al., 2016) and are sensitive indicators of desertification (Belnap et al., 2001; Coe et al., 2012; Read et

al., 2014; Delgado-Baquerizo et al., 2016). During dry periods and droughts, biocrusts often provide the only biological form of soil cover (Eldridge and Greene, 1994). Biocrusts such as lichens, mosses, fungi, cyanobacteria, eukaryotic algae, and other heterotrophic bacteria, live within or on the very top of soil surface (Belnap et al., 2003a; Rozenstein et al., 2014). They find almost all over the world, in all climatic regions, mesic environments tropical and temperate deserts and in the polar regions of the globe (West, 1990; Eldridge and Greene, 1994; Prasse and Bornkamm, 2000; Karnieli et al., 2002; Qin and Ferrara, 2007; Buis et al., 2009). Biocrusts play several important roles in arid and semi-arid ecosystems, including atmospheric nitrogen-fixation, increasing soil fertility, promoting water infiltration, seedling germination, increasing soil stability, improving soil properties and reducing soil erosion by wind (Greene and Darnall, 1990; Eldridge and Greene, 1994; Evans and Johansen, 1999; Eldridge et al., 2000; Belnap, 2003b; Belnap 2006; Miller et al., 2011; Liu et al., 2013 and Xu et al., 2013). Some field and laboratory studies indicated that biocrusts enhance the content of soil elements such as Ca, Mg, K, Fe, N, P and S (e.g. Belnap and Harper, 1995), and soil temperature (Belnap et al., 2001). Evidences show that biocrusts play also an important role in soil formation, preventing water and wind erosion and increasing soil fertility by N and C fixation, (Evans and Ehleringer, 1993; Lange et al., 1994). They provide suitable conditions for the establishment of vascular plants (Eldridge and Greene, 1994; Belnap and Lange, 2013; Belnap, 2003c). According to Belnap et al. (2001) up to 70% of the N fixed by cyanobacteria and cyanolichens is released into the surrounding soil environment and is available for vascular plants. Despite several studies have been done in the field of biocrusts globally (Belnap 2006; Chamizo et al., 2012a; Bowker et al., 2013; Chen and Duan, 2015), there are relatively few studies on the influences of biocrusts on soil physical and chemical characteristics. In this research, we studied soil characteristics under biocrusts as well as the interaction between biocrusts and soil in two geomorphologic sites with aridisols and solonchak

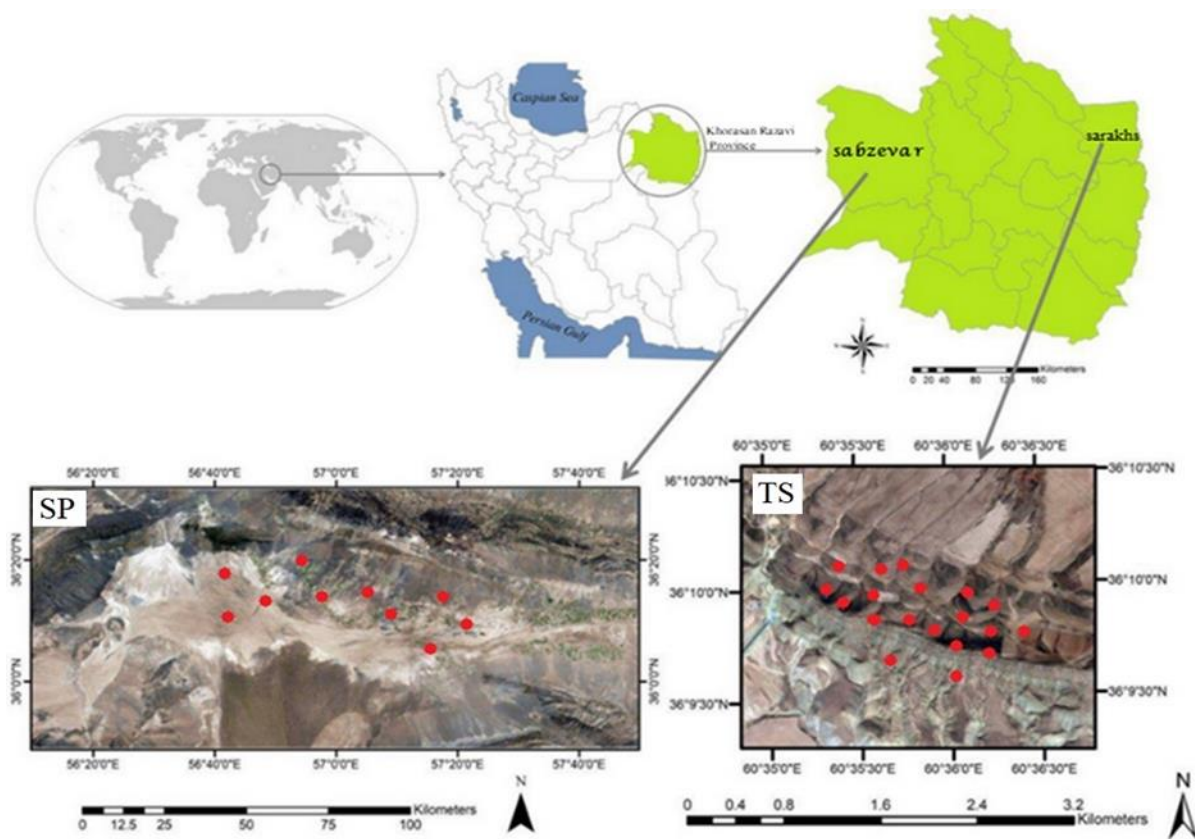
soils. It is hypothesized that biocrusts have positive effects on soil chemical and physical characteristics in the study area.

## 2. Data and Methods

### 2.1 Study sites

Around 75% of Iran is covered by drylands and regarding the land overuse and unsustainable land management, soil erosion is very intense in this area. Two sites in the semi-arid region of Khorasan-Razavi (north-eastern Iran) were selected (Fig. 1). Takhte Soltan (TS), with a size of 1854 km<sup>2</sup> is located at 36°16'00" – 36°15'00" N, 60°59'00" – 61°58'20"E. The topographical elevation of the study area varies between 750 m and 900 m a.s.l. It is a hilly area with a semi-arid (steppe) climate according to the Köppen climate classification. The average annual precipitation is 202 mm. January and February are the coldest months of the year with a mean air temperature of about 1° C. The highest average air temperature is 28.4° C in July. The dominant wind patterns include 120 days which usually blow from north and north-east, causing dust storms. The particle size distribution in the soil shows sandy loam textures. Rocks are mostly sandstone, conglomerate and shale (Table 1).

Sabzevar Playa (SP), with a size of about 2648 km<sup>2</sup> is located at 35° 55' 00" – 36° 25' 00" N, 56° 15' 00" – 57° 45' 00" E, in the eastern great Kavir basin. The topographical elevation of the study area varies between 750 m and 900 m a.s.l., The average topographical elevation of the region is over 750 m a.s.l. The general physiographic trend of the playa extends east to west along the Kal-Shour stream with an average length of 120 km. In addition to wind-borne sand dunes, volcanic rocks, tuffs and pelagic carbonate rocks formed in the upper Cretaceous period and earlier can be found around the playa. The study area has a semi-arid climate with annual precipitation of 150–200 mm and average annual temperature of 16–17° C. According to particle size classifications, soils of the investigation areas are generally sandy-loam in TS (Takhte Soltan) and silty in SP (Sabzevar Playa) (Table 1).



**Figure 1** Location of the study area in Khorasan Razavi province, Iran (sampling points are marked with red circles). TS (Takhte Soltan), SP (Sabzevar Playa).

**Table 1** Specifications of the study area.

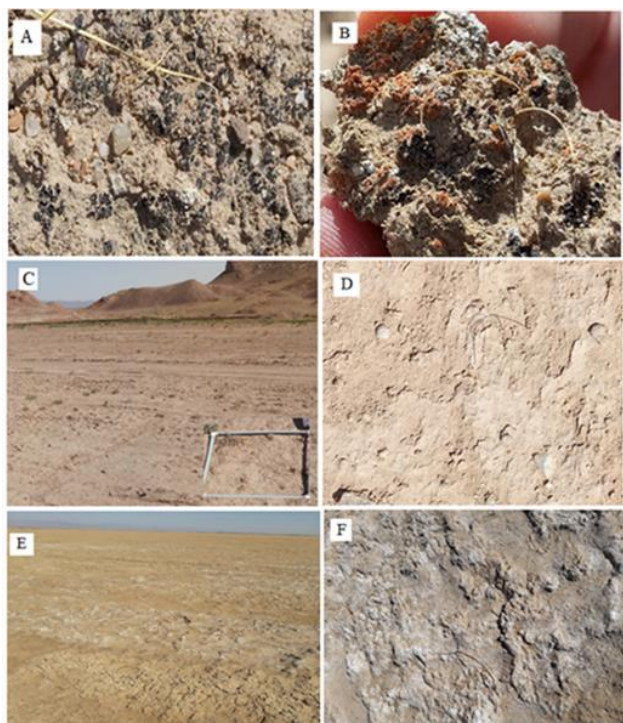
|                                 | SP                                                   | TS                                                   |
|---------------------------------|------------------------------------------------------|------------------------------------------------------|
| Climate (Koppen)                | BSK                                                  | BSK                                                  |
| Rainfall average (mm)           | 150–200                                              | 200–250                                              |
| Average annual temperature(° C) | 16–17                                                | 14–15                                                |
| Topographical elevation         | 650 m                                                | 750 m                                                |
| Soil classification             | Solonchak                                            | Aridisols–Vertisols                                  |
| Soil texture                    | silt                                                 | sandy loam                                           |
| Deposition                      | Cretaceous                                           | Quaternary–Cenozoic                                  |
| Geomorphology                   | Playa                                                | Pediment and Alluvial fans                           |
| Rock                            | sand dunes, volcanic rocks, tuffs, pelagic carbonate | sandstone, conglomerate, shale, gypsiferous mudstone |

## 2.2 Sampling

Samples were collected in September 2017 along a linear transect with a length of 50 m using 50 x 50 cm quadrates. All soil samples were taken in 10 replicates from the topsoil (upper 5 cm) as the purpose focused on Epilithic species living on topsoil and rocks. Distance between each quadrat was 5 meters

along the experimental transect according to its entire length. To compare the effects of biocrusts on the soil properties and vice versa, samples were collected in two zones including biocrusts and non-biocrusts. Twenty samples were taken from areas with biocrusts in the Takhte–Soltan region and areas

with non-BSC in the Sabzevar Playa region, respectively (Fig. 2). Samples were stockpiled in plastic bags and transferred to the laboratory of Natural Resources and Environment Faculty. The samples were kept at about 4° C in an isolated room.



**Figure 2** A and B biocrusts in TS (Takhte Soltan) ; C. landscape of the TS area; D. bare soil (without biocrusts) in TS; E. and F. bare soil in SP area; F. accumulation of minerals and salts on soil surface.

### 2.3 Laboratory analysis

The soil and biocrusts samples were air-dried and sieved to 2 mm. Sub-samples were mixed in a mechanical agate mortar to obtain the 0.5 mm particle size necessary for the determination of organic carbon and exchangeable cations. Soil samples were analysed for physicochemical characteristics as follows: pH was determined with a soil–water suspension with a ratio of 1: 2.5 by a pH meter (Rayment and Lyons, 2011); soil electrical conductivity (EC) was measured by a soil–water suspension in with a ratio of 1: 5 by an EC meter (Rayment and Lyons, 2011). Soil organic carbon (SOC) was estimated by the Walkley and Black method modified by Mingorance et al. (2007). Calcium carbonate (Nelson and Sommers, 1982), exchangeable cations (Ca, Mg, Na, K) and cation exchange capacity (CEC) were measured

by the formation of Cu (II) complexes with triethylenetetramine followed by photometric analyses (Meier and Kahr, 1999); soil moisture was calculated by the gravimetric (weighing) method.

In order to determine the cyanobacteria and green algae crusts, the soil surfaces were studied by stereomicroscope (Nikon SMZ–1) and optical microscope (Echo um–210 BD). Also, for better diagnosis and identification of species, samples were cultivated using the cultivation method applied by Kaushik (1987).

The study of moss and lichens was carried out through a morphological study by stereomicroscope (Nikon SMZ–1). Moreover, for the better evaluation of the lichens, the determination of the algae type and the details of the ascocarp-shaped structure have been done by sharp blade cutting and optical microscope observation (Echo um–210 BD). After cultivating, we identified species based on related researches including (Fang et al., 2007; Bilgrami and Saha, 2004; Komárek, 2005; Luton et al., 1999; Johnson et al., 2005; Hassler et al., 2012; Vijayan and Ray, 2015; Kumar et al., 2016; Moniri and Sipman, 2009; Zedda et al., 2000; McCune, 2007).

For scanning electron microscopy SEM, small part of the undisturbed samples was glued onto Aluminium stubs with the exposed vertical natural section facing upward and coated gold or platinum (Cox et al., 1989). Observations of the microstructures of soil crust were performed on a Leo 1450VP Scanning Electron Microscope (SEM, LEO, and Germany).

### 2.4 Statistical analysis

The effects of biocrusts on soil physicochemical properties on soil and differences between biocrusts, non-biocrusts and bare soils were analyzed in SPSS (Version 20.0) with t-test ( $P < 0.05$ ) used to test. At first, the parameters were normalized by the Kolmogorov–Smirnov test. Then, by using the t-test, in the 5% statistical significance, the groups were evaluated. To investigate the effect of biocrusts on physicochemical parameters of the soil were compared in each catchment (non-biocrusts layer was compared with biocrusts surfaces).

### 3. Results

#### 3.1 Biocrust influences on soil physicochemical properties

The current study indicated that biocrusts have significant effects on most physical and chemical characteristics (Table 2). Na, K, Ca, and Mg content, EC and SOC have changed significantly between different sites.

Sand, silt and clay contents varied significantly under the crust types in TS (Table 2). No significant difference in soil texture was found between the biocrusts and non-biocrusts layers. A significant difference in sand and silt contents was observed between soils with a crust (biocrusts and non-biocrusts) and bare soils. Although clay content of bare soil was higher, the difference was statistically significant. Comparison of sand, silt, and clay contents between the biocrusts site and the non-biocrusts site show significant differences for all three-particle size classes.

Soil texture by affecting on soil moisture, play an important role in composition and distribution of biocrusts (Kidron, 2016), the dominant texture is sandy-loam in TS and silty in SP. The amount of sand under biocrusts was significantly higher than in bare soils, and in a non-biocrusts site, the percentage of clay was significantly higher. Soil moisture content was higher in soils without biocrusts in surface layer, but was not statistically significant. In the TS site, the biocrusts samples and the non-biocrusts samples did not differ significantly regarding moisture. In soil with biocrusts the electrical conductivity (ds/m) was significantly lower than bare soils.

Regarding soil pH, there were significant differences ( $p < 0.05$ ) between pH under biocrusts and bare soil (Table 2). Moreover, soil pH under biocrusts and non-biocrusts site was not statistically significant. Organic carbon in biocrusts soils was lower than bare soils, according to Table 2. The total content of soil organic carbon in the upper 5 cm of the soils under non-biocrusts was significantly lower than the organic carbon content under biocrusts. The amount of calcium carbonate in the TS region (biocrusts area) was higher than SP (bare soil), but this difference was not statistically significant. For Na, Ca, Mg and K exchange, there is a significant difference between the two sites, the number of ex-

changeable cations for the bare soil area was significantly higher. Exchangeable cations content shows no variation between soil under biocrusts and non-biocrusts.

The cation exchange capacity, SAR and ESP of soils are more dependent on the type of soil and soil organic carbon content. As a result, due to the high levels of above two factors, the CEC, SAR, and ESP are also significantly higher in the bare soil site. Exchangeable cation content, CEC, SAR, and ESP show no significant differences between soil under biocrusts and non-biocrusts.

Micrograph of the samples in ppl (plane-polarized light) and scanning electron microscopy (SEM) observations of biocrusts in TS showed aggregations of fungal hyphae (Fig. 4). Connecting of individual grains was observed by the hyphae of lichens and fungi, polysaccharide of cyanobacteria (Fig. 4).

#### 3.2 The dominant species of Biocrusts in the TS region

Biocrusts site presented higher cyanobacteria, green algae and lichens biodiversity and species richness than non-biocrusts site (Figs. 7 and 8; Table 4). The dominant species of cyanobacteria include: *Oscillatoria limosa*, *Oscillatoria irrigua*, *Phormidium favosum*, *Phormidium ambiguum*, *Phormidium uncinatum*, *Pseudanabaena sp*, and *Microcoleus vaginatus*.

Some of these species such as *Leptolyngbya boryana*, *Microcoleus vaginatus*, *Nostoc commune*, *Oscillatoria curvipes* and *Oscillatoria tenuis* were also found in the non-biocrusts soil. The dominant species of green algae including: *Neochloris sp*, *Chlorococcum sp*, *Chlorella sp*, *Planktosphaeria*, *Rhizodolium*, *Oedogonium*, *Microspora*, *Euglena* and *Protococcus* observed on both sites with biocrusts and non-biocrusts. Two species *Chlorococcum sp* and *Chlorella sp* were also observed in the soil under abandoned agricultural land. The dominant lichens also in the study area are: *Collema tenax*, *Psora decipiens*, *Toninia candida*, *Caloplaca microthallina*, *Circinaria mansourii*, *Leptogium*, *Candelariella vitellina*, *Gyalolechia subbracteata*, *Rinodina bischoffii*, *Endocarpon unifoliatum*, *Endocarpon pusillum* and *Sarcogyne*.

**Table 2** Statistical assessment of mean and variance differences between two sites TS (biocrusts) and SP (bare soils).

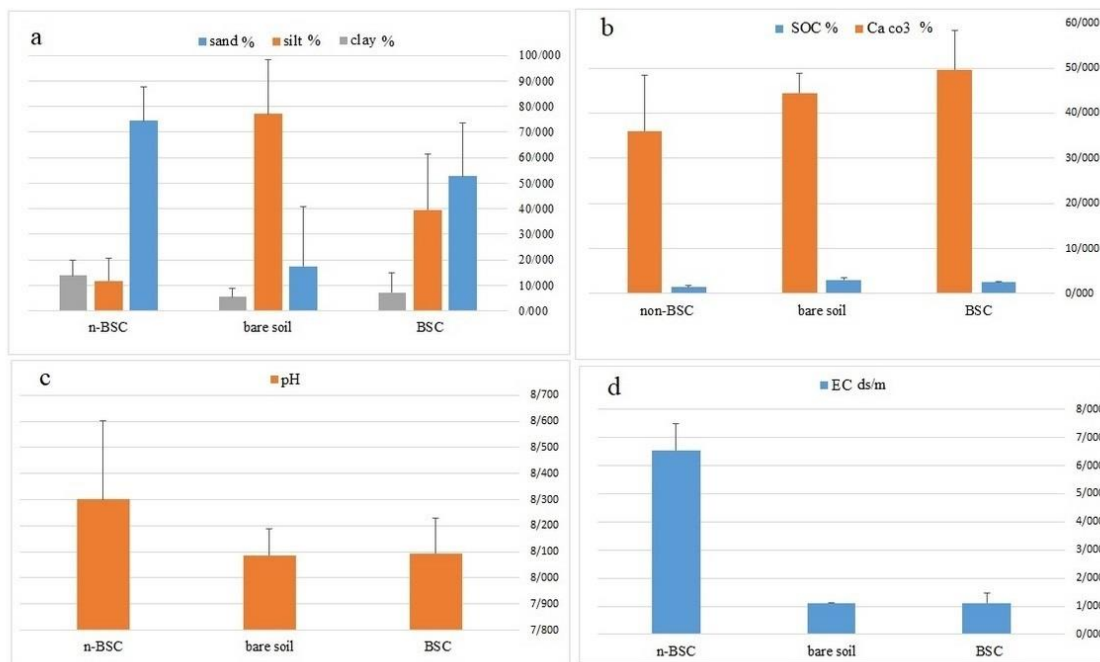
| Elements          | Levene's Test for Equality of Variances |       | t-test for Equality of Means |        |                 | Mean Difference | Std. Error Difference |
|-------------------|-----------------------------------------|-------|------------------------------|--------|-----------------|-----------------|-----------------------|
|                   | F                                       | Sig.  | t                            | df     | Sig. (2-tailed) |                 |                       |
| EC ds/m*          | 26.652                                  | 0.000 | -17.295                      | 9.000  | 0.000           | -1.923          | 0.111                 |
| pH                | 1.355                                   | 0.260 | 0.158                        | 18.000 | 0.876           | 0.008           | 0.054                 |
| SOC %*            | 0.592                                   | 0.452 | -3.513                       | 18.000 | 0.002           | -0.567          | 0.161                 |
| CaCo <sub>3</sub> | 7.365                                   | 0.014 | 1.704                        | 13.342 | 0.112           | 5.2041          | 3.05                  |
| Na (mg/kg)*       | 0.192                                   | 0.667 | -5.367                       | 18.000 | 0.000           | -1.601          | 0.298                 |
| K (mg/kg)*        | 0.014                                   | 0.908 | -7.299                       | 18.000 | 0.000           | -28.254         | 3.871                 |
| Ca (mg/kg)*       | 3.890                                   | 0.064 | -9.314                       | 18.000 | 0.000           | -1.943          | 0.209                 |
| Mg (mg/kg)*       | 0.276                                   | 0.606 | -2.449                       | 18.000 | 0.025           | -0.951          | 0.388                 |
| SAR*              | 8.353                                   | 0.010 | -12.041                      | 9.085  | 0.000           | -1.911          | 0.159                 |
| ESP*              | 14.922                                  | 0.001 | -12.399                      | 9.083  | 0.000           | -1.812          | 0.146                 |
| CEC*              | 1.899                                   | 0.185 | -3.113                       | 18.000 | 0.006           | -1.051          | 0.338                 |
| Moisture%         | 0.013                                   | 0.911 | -0.739                       | 18.000 | 0.469           | -0.073          | 0.098                 |
| sand*             | 0.414                                   | 0.528 | 3.573                        | 18.000 | 0.002           | 35.235          | 9.862                 |
| silt*             | 0.005                                   | 0.942 | -3.915                       | 18.000 | 0.001           | -37.657         | 9.619                 |
| clay*             | 4.363                                   | 0.051 | 0.609                        | 18.000 | 0.550           | 1.647           | 2.706                 |

\*represent significant at 95% confidence interval or P&lt;0.05

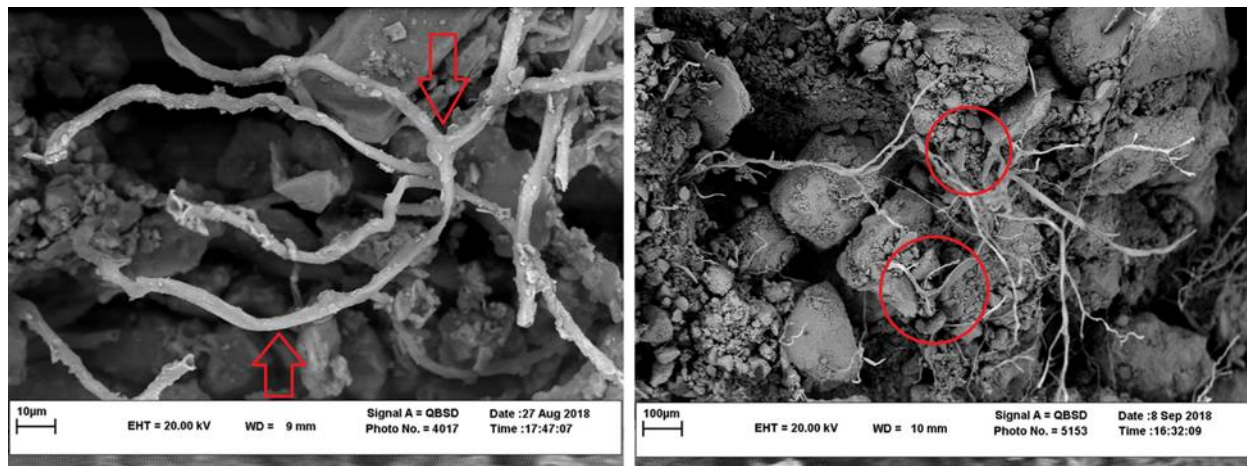
**Table 3** Statistical assessment of mean and variance differences between two sites biocrusts and non-biocrusts in TS area.

| Elements          | Levene's Test for Equality of Variances |       | t-test for Equality of Means |        |                 | Mean Difference | Std. Error Difference |
|-------------------|-----------------------------------------|-------|------------------------------|--------|-----------------|-----------------|-----------------------|
|                   | F                                       | Sig.  | t                            | df     | Sig. (2-tailed) |                 |                       |
| EC ds/m           | 10.460                                  | 0.005 | -1.539                       | 11.477 | 0.151           | -0.488          | 0.317                 |
| pH                | 4.276                                   | 0.053 | -1.974                       | 18.000 | 0.064           | -0.207          | 0.105                 |
| SOC %*            | 2.156                                   | 0.159 | 5.888                        | 18.000 | 0.000           | 1.025           | 0.174                 |
| CaCo <sub>3</sub> | 1.194                                   | 0.289 | 2.847                        | 18.000 | 0.011           | 13.577          | 4.769                 |
| Na (mg/kg)        | 0.014                                   | 0.908 | 0.299                        | 18.000 | 0.768           | 0.082           | 0.274                 |
| K (mg/kg)         | 0.285                                   | 0.600 | 0.501                        | 18.000 | 0.622           | 2.080           | 4.149                 |
| Ca (mg/kg)        | 0.283                                   | 0.601 | -1.187                       | 18.000 | 0.251           | -0.326          | 0.275                 |
| Mg (mg/kg)        | 0.006                                   | 0.941 | 1.145                        | 18.000 | 0.267           | 0.396           | 0.346                 |
| SAR               | 0.203                                   | 0.658 | -0.339                       | 18.000 | 0.739           | -0.085          | 0.251                 |
| ESP               | 0.569                                   | 0.460 | 0.202                        | 18.000 | 0.842           | 0.054           | 0.267                 |
| CEC               | 0.127                                   | 0.725 | -1.194                       | 18.000 | 0.248           | -0.600          | 0.502                 |
| Moisture%         | 5.278                                   | 0.034 | -1.136                       | 12.434 | 0.277           | -0.194          | 0.171                 |
| sand*             | 1.944                                   | 0.180 | -2.792                       | 18.000 | 0.012           | -21.773         | 7.798                 |
| silt*             | 6.408                                   | 0.021 | 3.728                        | 11.909 | 0.003           | 27.704          | 7.431                 |
| clay*             | 0.156                                   | 0.697 | -2.127                       | 18.000 | 0.047           | -6.704          | 3.152                 |

\*represent significant at 95% confidence interval or P&lt;0.05



**Figure 3** Mean ( $\pm$ SD, n=30) sand, silt and clay in the three soil layers (a) Calcium carbonate (CaCO<sub>3</sub>) and Soil organic carbon content (SOC) (b) pH (c) and EC ds/m (d). (BSC and n-BSC) from TS region.



**Figure 4** The polysaccharide attached Clay particles. The polysaccharide filaments that enclose the clay particles are marked in the figure with arrows and a red circle.

## 4. Discussion

Biocrusts affect nutrient cycling as well as hydrological processes in early soil succession processes. Several studies show that biocrusts could affect the weathering of rocks and thus contribute to the formation of soil and distribution of the particle size (Souza-Egipsy et al., 2004; Aghamiri and Schwartzman, 2002). Moreover, many studies have indicated an increase in soil moisture (Issa et al., 2007), N and

SOC content in the soils with biocrusts (Rogers and Burns, 1994; Gao et al., 2010).

### 4.1 The effect of biocrusts on physicochemical properties

In our study, some physicochemical properties such as EC, pH, SOC, exchangeable cations, SAR, ESP, CEC and moisture were lower in biocrusts compared to bare soils.

Soil pH under the biocrusts areas was lower than bare and non-biological crust soils, although there is not significant difference between sites. Increase in microbial population, and carbon dioxide generation as well as higher respiration found in biocrusts could decrease pH (Büdel, 2005; Lane et al., 2013). Chamizo et al. (2012b) and Miralles et al. (2012) found that the soil pH decreased with the development of biocrusts from primary sequences such as cyanobacteria to the final sequence, such as mosses and lichens. Miralles et al. (2013) found that compared to the soil without crusts with cyanobacterium and lichen substrates, pH for coverless soil <cyanobacteria> lichen, which indicates a decrease in soil pH in the presence and development of biocrusts. There is an interaction between soil pH and soil organic carbon. Biocrusts have a significant role in producing organic carbon through the carbon fixation (Beymer and Cleopatek, 1991; Danin and Ganor, 1991). Phillips and Belnap (1998) indicated that carbon stabilization increases in the presence of licorice and moss. Increasing in the infiltration and reduction of cations at the surface and consequently the decrease of pH at the soil level is accurate for comparison between two bare and crustal zones, but the amount of cations in two sites with crust and without crust is not significant and the lower pH content in the soil to the crusty surface can be associated with increased microbial respiration (Chamizo et al., 2012a).

In this study, bare soils (playa) had much higher values of sodic and saline properties such as Ca, Na, Mg, EC, ESP and SAR. Similarly, Abed et al. (2013) showed that biocrusts soils had less pH and EC than bare soils. Subsequent evaporation and ponding of water in these areas would leave substantial amounts of salts behind (Kakeh et al., 2018). High evaporation causes salinity that could build up to the surface (Fig. 2F). Increasing salinity has reduced colonization of plants or biocrusts, thus also increasing the salinity trend (Kakeh et al., 2018). The amount of sodium and potassium in the non-biocrusts compared with biocrusts has increased, which confirms the studies of Maqubela et al. (2009), which explain the reason for the increase of some soil elements due to the presence of biocrusts, as follows: copper, sodium, potassium, and

zinc stick to the outer surface of the cell wall of the mosses. When wet and does dry, the elements are washed from the lichens and enter the soil, due to their positive load they are absorbed by clay colloids that have negative charges, resulting in an increase in their amount in such soils. Significant changes in the soil EC are observed between the biocrusts and bare soils (Table 2).

Increasing salinity and clay content prevents vegetation deployment and biocrusts (increasing clay content in non-biocrusts). Increasing clay accumulation at the soil surface decreases soil infiltration. It will only be able to moisturize the superficial layer, in this situation, an increase in the amount of evaporation from the surface layer of the soil leads to an increase in the accumulation of minerals from the elements of sodium, calcium and magnesium in this layer. So, the high amount of calcium and magnesium in the non-biocrusts and bare soils is justified.

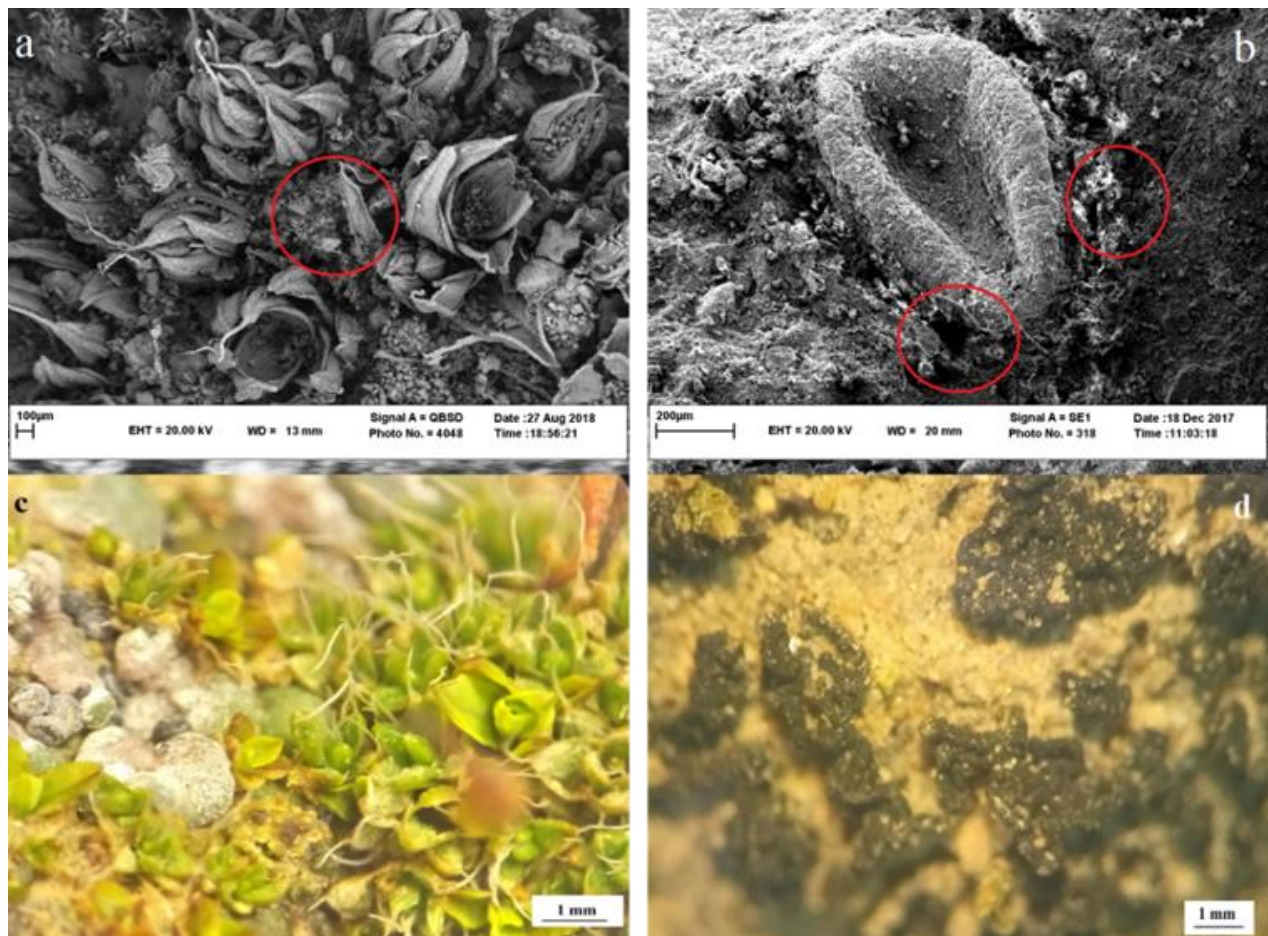
Many studies have shown the effective role of biocrusts in the weathering of parent material, soil formation and particle size distribution (Aghamiri and Schwartzman, 2002; Souza-Egipsy et al., 2004). Soil organic carbon was higher in biocrusts surface soils than non-biocrusts, likely due to the ability of biocrusts to fix atmospheric C (Thomas et al., 2008; Zaady et al., 2000). This leads to enhanced biocrust biomass and thus SOC in the surface soil layer (0–5cm) (Chamizo et al., 2012a). Even a thin layer of photo synthetically layer on the soil can be important sources of organic carbon input (Zaady et al., 2000). Biocrusts also produce and secrete extracellular polysaccharides into surrounding soils, increasing the soil C pool (Mager and Thomas, 2011).

Biocrusts have an important role in the production of the SOC by carbon fixation (Beymer and Klopatek, 1991) and decomposition of organic matter (Danin and Ganor, 1991) in arid soils. Phillips and Belnap (1998) and Zhao et al. (2013) argued that carbon stabilization has been increased in the presence of lichen and moss. Base structures have more organic carbon compared to other structures, and there is an inverse relationship between SOC and soil pH. An increase in carbon in the Playa region (non-biocrusts) is due to the dust that is deposited by the wind and the habitat of ants and termites

(Kidron, 2016). Cyanobacteria can induce carbon dioxide to calcium carbonate (Wierchos et al., 2012); because of the decrease in calcium carbonate content in non-biocrusts and bare soils, it can be concluded that cyanobacteria decrease at these levels. Soil texture shows the relative frequency of sand, clay and silt particles in soil samples.

Previous study has shown that soils with silty loamy texture are more likely to hold different populations of cyanobacteria, lichens and mosses compared to clay soils (Kleiner and Harper, 1997). Hydrological processes, aggregation and soil for-

mation can be related to biological activities (Miralles et al., 2012). Sediment and desert dusts can aggregate with wind velocity, so in small catchments, the aggregation will be done in rock cracks (Hirmas et al., 2011). In a wider range, it improves topography and sediment accumulation patterns, the solid rocky surface of the desert pavement, with a smooth topography, capture less dust (Hirmas et al., 2011). As a result, a significant increase in the percentage of clay in the biocrusts compared with the bare soil can be attributed to dust capture in the biocrusts of the soil surface.



**Figure 5** The red circle in image shows that fungal hyphae are attached to the soil. a) SEM of mosses; b) SEM of lichens; c) *Tortula atrovirens*; d) *Collema sp.*

#### 4.2 The role of biocrusts in soil stability

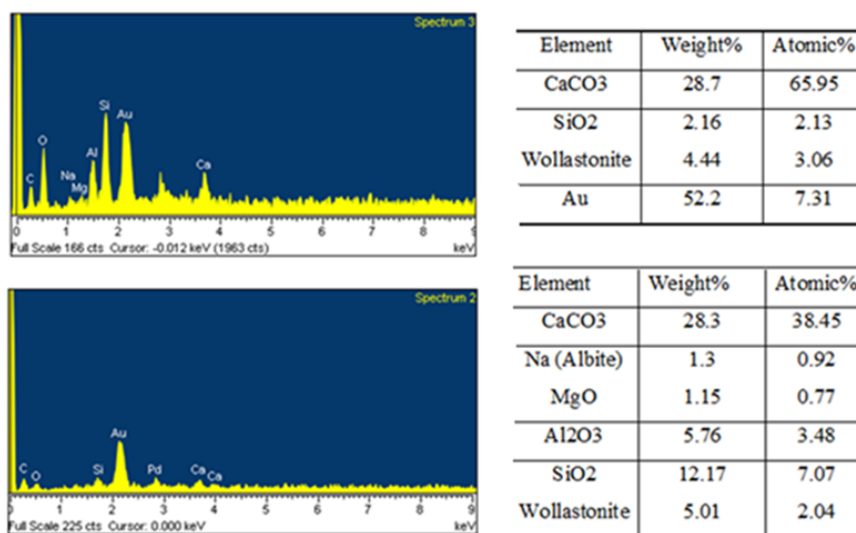
The biological activities of cyanobacteria, mosses and lichens have an important impact on soil erosion. The biocrust formed an enhanced contact structure composed of fine mineral particles. Early

cyanobacterial crusts become stabilized when cyanobacteria polysaccharides that bind the soil particles, whereas late cyanobacterial crusts achieve stabilization by secreting filamentous and polysaccharides (Zhang et al., 2016). Mosses trap dust among

their caulidia and fix by rhizoid (Figs. 4 and 5). The extensive connections between sand grains of the sheath material can be seen, and multiple sheets can often be seen attached to the same sand grain. The polysaccharide attached clay particles. This may be a mechanism by which cyanobacteria increase mineral availability to themselves and vascular plants. Positively charged macronutrients are bound to both negatively charged clay particles and the sheath materials; thus, they are held in the upper soil horizons in a form readily available to vascular

plants, instead of being leached away (Belnap and Gardner, 1993).

Cyanobacteria and lichens fix dust particles to their polysaccharide sheaths, which become sticky when wet (Farpoor and Krouse, 2008). Trapped dust was thought to be important for creating a crust rich in clay and fine silt (Fig. 4). At the lichen stage, the crusts bind sand particles tightly with mycelium to form a stable layer. The moss profile shows that the moss rhizomes have kept the particles of soil building an extremely stable surface (Fig. 5).



**Figure 6** The EDS spectrum of the soil surface, along with the lichens and mosses biofilm and the percentage of its elements

Recovery or development of biocrusts makes changes on soil stability, resulting in increased biological activity. Consistent with our study, Danin and Ganor (1991) found calcite concentrations increased in biological surface soil. Hu et al. (2002) explained that changes in the numbers and species of algae were responsible for the mineralogical component variety in bio-crust at Shapoutou. Hyphae of lichen and rhizoid of mosses can physically break down quartz. Lamas et al. (1995) found that the influence and growth of hyphae occur both horizontally and vertically and may exceed 5 mm, mainly via intergranular voids. Mineral surfaces are often affected by considerable amounts of organic covering (Krumbein and Dyer, 1985) that are produced by cyanobacteria, fungi and algae as well as higher organisms that inhabit external and internal (fissure/fracture) surfaces of mineral substrates (Fig. 4; polysaccharides and the eroding effect).

This research showed that biocrusts contribute to mineral erosion, and after analysing the physico-chemical properties, we found that the main properties which showed significant differences with crust and bare soils were SOC, CaCO<sub>3</sub> and amount of clay.

## 5. Conclusion

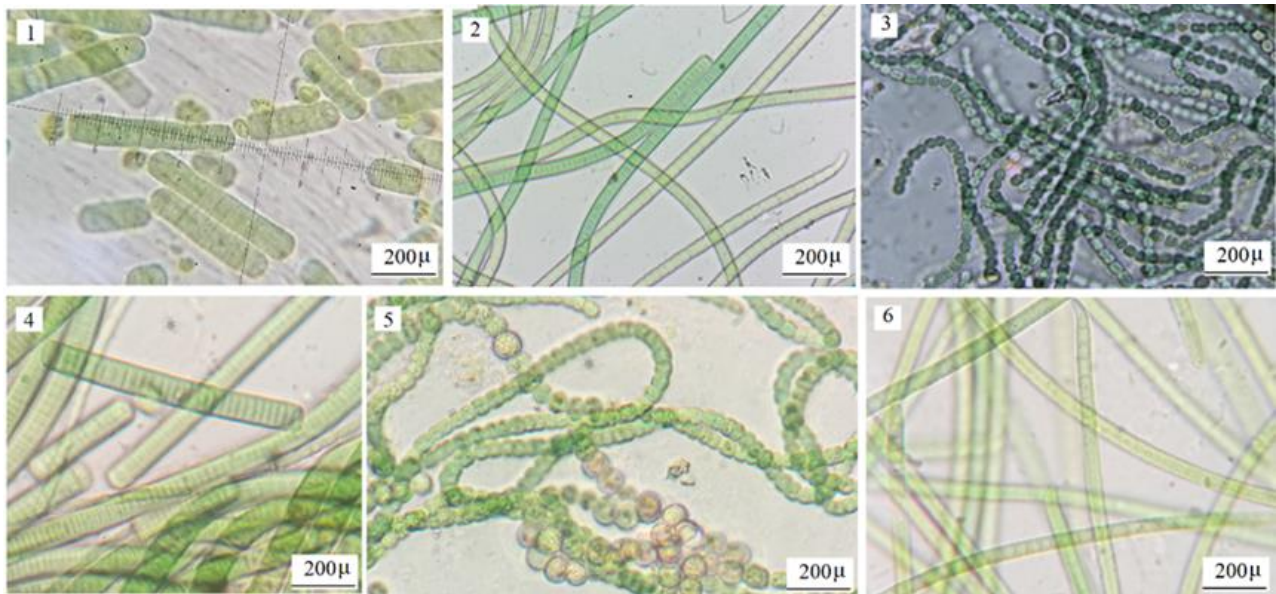
Cyanobacteria can grow on the soil surface hence promoting the formation of biocrusts in arid and semi-arid areas and protect the soil against erosion conditions. Experimental analyses indicate that key physicochemical properties of OC and CaCO<sub>3</sub> differed between biocrusts and bare soil significantly in terms of different types of biocrusts. In general, in the TS, physicochemical property measurements were higher in bio-crusted surface soils when compared to the bare soil at the same depth. It is con-

cluded that in the semiarid catchment in eastern Iran the presence of biocrusts generally increased the positive qualities of the soil. Our study reinforces the notion that biocrusts are important for maintaining and improving the soil conditions.

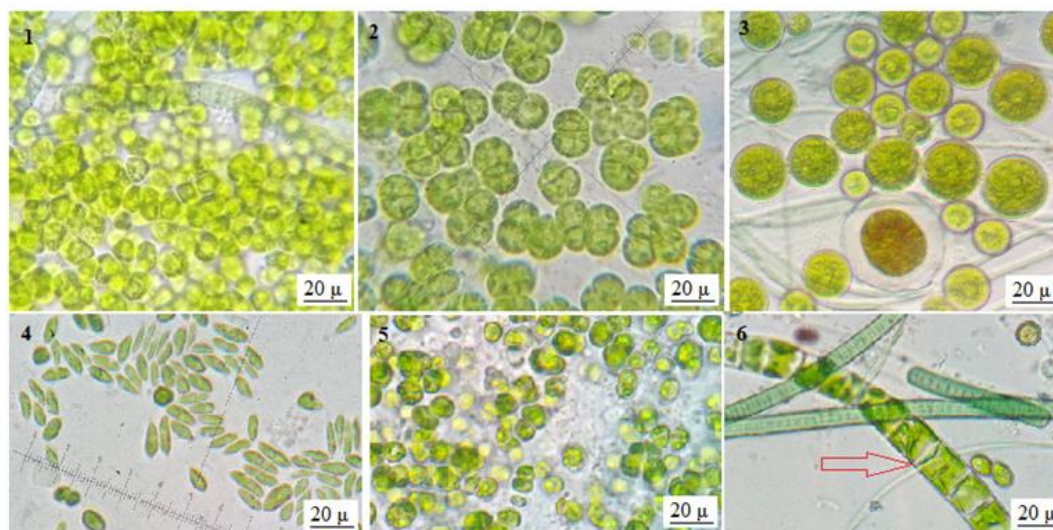
### Acknowledgements

This work was funded by "Ferdowsi University of Mashhad" (Grant Number: 3-44201).

### Appendix



**Figure 7** Cyanobacteria species identified in soil surface samples (0–5 cm) at the TS region: 1) *Leptolyngbya* sp.; 2) *Phormidium* sp.; 3) *Komvophoron* sp.; 4) *Oscillatoria* sp.; 5) *Nostoc* sp.; 6) *Phormidium* sp.



**Figure 8** Green algae species identified in soil surface samples (0–5 cm) at the TS region: 1) *Chlorella* sp.; 2) *Protococcus* sp.; 3) *Chlorococcum* sp.; 4) *Euglena* sp.; 5) *Planktosphaeria* sp.; 6) *Microspora* sp.

**Table 4** Biocrusts species found in samples of TS region and playa. Detailed picture of the different cyanobacteria and green algae species are shown in Figures 7 and 8.

|                               | Cyanobacteria |          |       |                        | Green Algae |          |       |                                 | Lichen |          |       |                              | Bryophyte |          |       |
|-------------------------------|---------------|----------|-------|------------------------|-------------|----------|-------|---------------------------------|--------|----------|-------|------------------------------|-----------|----------|-------|
|                               | BSCs          | NON-BSCs | Playa |                        | BSCs        | NON-BSCs | Playa |                                 | BSCs   | NON-BSCs | Playa |                              | BSCs      | NON-BSCs | Playa |
| <i>Leptolyngbya boryana</i>   | ✓             | ✓        | -     | <i>Neochloris</i>      | ✓           | ✓        | -     | <i>Collema tenax</i>            | ✓      | -        | -     | <i>Bryum argenteum</i>       | ✓         | -        | -     |
| <i>Microcoleus vaginatus</i>  | ✓             | ✓        | -     | <i>Chlorococcum</i>    | ✓           | ✓        | ✓     | <i>Psora decipiens</i>          | ✓      | -        | -     | <i>Tortula atrovirens</i>    | ✓         | -        | -     |
| <i>Nostoc commune</i>         | ✓             | ✓        | -     | <i>Chlorella</i>       | ✓           | ✓        | ✓     | <i>Toninia candida</i>          | ✓      | -        | -     | <i>Syntrichia ruralis</i>    | ✓         | -        | -     |
| <i>Oscillatoria curviceps</i> | ✓             | ✓        | -     | <i>Planktosphaeria</i> | ✓           | ✓        | -     | <i>Caloplaca microthallina</i>  | ✓      | -        | -     | <i>Syntrichia caninervis</i> | ✓         | -        | -     |
| <i>Oscillatoria tenuis</i>    | ✓             | ✓        | -     | <i>Rhizoclonium</i>    | ✓           | ✓        | -     | <i>Circinaria mansourii</i>     | ✓      | -        | -     |                              |           |          |       |
| <i>Oscillatoria chalybea</i>  | ✓             | -        | -     | <i>Oedogonium</i>      | ✓           | -        | -     | <i>Leptogium</i>                | ✓      | -        | -     |                              |           |          |       |
| <i>Oscillatoria limosa</i>    | ✓             | -        | -     | <i>Protococcus</i>     | ✓           | ✓        | -     | <i>Candelariella vitellina</i>  | ✓      | -        | -     |                              |           |          |       |
| <i>Oscillatoria irrigua</i>   | ✓             | -        | -     | <i>Microspora</i>      | ✓           | -        | -     | <i>Caloplaca microthallina</i>  | ✓      | -        | -     |                              |           |          |       |
| <i>Phormidium favosum</i>     | ✓             | -        | -     | <i>Euglena</i>         | ✓           | -        | -     | <i>Caloplaca microthallina</i>  | ✓      | -        | -     |                              |           |          |       |
| <i>Phormidium amabilegum</i>  | ✓             | -        | -     |                        |             |          |       | <i>Gyalolechia subbracteata</i> | ✓      | -        | -     |                              |           |          |       |
| <i>Phormidium uncinatum</i>   | ✓             | ✓        | -     |                        |             |          |       | <i>Rinodina bischoffi</i>       | ✓      | -        | -     |                              |           |          |       |
| <i>Pseudanabaena</i>          | ✓             | -        | -     |                        |             |          |       | <i>Endocarpon unifoliatum</i>   | ✓      | -        | -     |                              |           |          |       |
|                               |               |          |       |                        |             |          |       | <i>Endocarpon pusillum</i>      | ✓      | -        | -     |                              |           |          |       |
|                               |               |          |       |                        |             |          |       | <i>Sarcogyne</i>                | ✓      | -        | -     |                              |           |          |       |

## References

- Abed RM, Al-Sadi AM, Al-Shehi M, Al-Hinai S, Robinson MD. 2013. Diversity of free-living and lichenized fungal communities in biological soil crusts of the Sultanate of Oman and their role in improving soil properties. *Soil Biology and Biochemistry*, **57**: 695–705.
- Aghamiri RR, Schwartzman DW. 2002. Weathering rates of bedrock by lichens: a mini watershed study. *Chemical Geology*, **188**(3–4): 249–259.
- Belnap J. 2003(a). The world at your feet: desert biological soil crusts. *Frontiers in Ecology and the Environment*, **1**(4): 181–189.
- Belnap J. 2003 (b). Factors influencing nitrogen fixation and nitrogen release in biological soil crusts. In: Belnap J, Lange O. (eds.): *Biological soil crusts: structure, function, and management*, Springer-Verlag, Berlin, 241–261.
- Belnap J. 2003 (c). Biological soil crusts in deserts: a short review of their role in soil fertility, stabilization, and water relations. *Algological Studies*, **109**(1): 113–126.
- Belnap J. 2006. The potential roles of biological soil crusts in dryland hydrologic cycles. *Hydrological Processes: An International Journal*, **20**(15): 3159–3178.
- Belnap J, Büdel B, Lange OL. 2001. Biological soil crusts: characteristics and distribution. In: *Biological soil crusts: structure, function, and management*, Springer, Berlin, Heidelberg, 3–30.
- Belnap J, Büdel B, Lange OL. 2003. Biological soil crusts: characteristics and distribution. In: Belnap J, Lange OL (eds.): *Biological soil crusts: structure, function and management*, Springer-Verlag: Berlin; 3–30.
- Belnap J, Gardner JS. 1993. Soil microstructure in soils of the Colorado Plateau: the role of the cyanobacterium *Microcoleus vaginatus*. *The Great Basin Naturalist*, 40–47.
- Belnap J, Harper KT. 1995. Influence of cryptobiotic soil crusts on elemental content of tissue of two desert seed plants. *Arid Soil Res Rehabil*, **9**: 107–115.
- Belnap J, Lange OL. 2013. Eds.: *Biological soil crusts: structure, function, and management*. Springer Science & Business Media (Vol **150**).
- Beymer R, Klopatek J. 1991. Potential contribution of carbon by microphytic crusts in pinyon-juniper woodlands. *Arid Land Research and Management*, **5**(3): 187–198.
- Bilgrami KS, Saha LC. 2004. *A Textbook of Algae*. CBS Publishers and Distributors, 263p.
- Bowker MA, Eldridge DJ, Val J, Soliveres S. 2013. Hydrology in a patterned landscape is co-engineered by soil-disturbing animals and biological crusts. *Soil Biol. Biochem.*, **61**(suppl. C): 14–22.
- Büdel B. 2005. Microorganisms of biological crusts on soil surfaces. In: In: 425 Varma, A., Buscot, F. (Eds.), *Microorganisms in Soils: Roles in Genesis and Functions*, vol. 3, Springer, Berlin, pp. 307–323.
- Buis CI, Geuken E, Visser DS, Kuipers F, Haagsma EB, Verkade HJ, Porte RJ. 2009. Altered bile composition after liver transplantation is associated with the development of nonanastomotic biliary structures. *Journal of hepatology*, **50**(1): 69–79.
- Chamizo S, Cantón Y, Miralles I, Domingo F. 2012a. Biological soil crust development affects physicochemical characteristics of soil surface in semiarid ecosystems. *Soil Biol. Biochem.*, **49**: 96–105.
- Chamizo S, Cantón Y, Lázaro R, Solé-Benet A, Domingo F. 2012b. Crust composition and disturbance drive infiltration through biological soil crusts in semiarid ecosystems. *Ecosystems*, **15**: 148–161.
- Chen X, Duan Z. 2015. Impacts of soil crusts on soil physicochemical characteristics in different rainfall zones of the arid and semi-arid desert regions of northern China. *Environmental Earth Sciences*, **73**(7): 3335–3347.
- Coe BP, O’Roak BJ, Vives L, Girirajan S, Karakoc E, Krumm N, Levy R, Ko A, Lee C, Smith JD, Turner EH. 2012. Sporadic autism exomes reveal a highly interconnected protein network of de novo mutations. *Nature*, **485**(7397): 246–250.
- Cox AC, Hukins DWL, Sutton TM. 1989. Infection of catheterised patients: bacterial colonisation of encrusted Foley catheters shown by scanning electron microscopy. *Urological research*, **17**(6): 349–352.
- Danin A, Ganor E. 1991. Trapping of airborne dust by mosses in the Negev Desert, Israel. *Earth Surface Processes and Landforms*, **16**(2): 153–162.
- Delgado-Baquerizo M, Maestre FT, Bowker MA, Eldridge DJ, Cortina J, Lázaro R, Gallardo A, Berdugo M, Castillo-Monroy AP, Valencia E. 2016. Biological soil crusts as a model system in ecology. In: *Biological Soil Crusts: An Organizing Principle in Drylands*, Springer, Cham., 407–425.
- Eldridge DJ, Greene RSB. 1994. Assessment of sediment yield by splash erosion on a semi-arid soil with varying cryptogam cover. *Journal of Arid Environments*, **26**: 221–232.
- Eldridge DJ, Zaady E, Shachak M. 2000. Infiltration through three contrasting biological soil crusts in patterned landscapes in the Negev, Israel. *Catena*, **40**(3): 323–336.
- Evans RD, Ehleringer JR. 1993. A break in the nitrogen cycle in aridlands. Evidence from  $\delta^{15}\text{N}$  of soils. *Oecologia*, **94**: 314–317.
- Evans RD, Johansen JR. 1999. Microbiotic crusts and ecosystem processes. *Critical reviews in plant sciences*, **18**(2): 183–225.

- Fang HY, Cai QG, Chen H, Li QY. 2007. Mechanism of formation of physical soil crust in desert soils treated with straw checkerboards. *Soil Tillage Res.*, **93**: 222–230.
- Farpoor MH, Krouse HR. 2008. Stable isotope geochemistry of sulfur bearing minerals and clay mineralogy of some soils and sediments in Loot Desert, central Iran. *Geoderma*, **146**(1): 283–290.
- Gao S, Ye X, Chu Y, Dong M. 2010. Effects of biological soil crusts on profile distribution of soil water, organic carbon and total nitrogen in Mu Us Sandland, China. *Journal of Plant Ecology*, **3**: 279–284.
- Greene B, Darnall DW. 1990. Microbial oxygenic photoautotrophs (cyanobacteria and algae) for metal-ion binding. *Microbial mineral recovery*, 277–302.
- Hirmas DR, Graham RC, Kendrick KJ. 2011. Soil-geomorphic significance of land surface characteristics in an arid mountain range, Mojave Desert, USA. *Catena*, **87**(3): 408–420.
- Hu C, Liu Y, Song L, Zhang D. 2002. Effect of desert soil algae on the stabilization of fine sands. *Journal of Applied Phycology*, **14**: 281–292.
- Issa OM, Défarge C, Le Bissonnais Y, Marin B, Duval O, Bruand A, d'Acqui LP, Nordenberg S, Annerman M. 2007. Effects of the inoculation of cyanobacteria on the microstructure and the structural stability of a tropical soil. *Plant and soil*, **290**(1–2): 209–219.
- Johnson SL, Budinoff CR, Belna J, Garcia-Pichel F. 2005. Relevance of ammonium oxidation within biological soil crust communities. *Environmental Microbiology*, **7**(1): 1–12.
- Kakeh J, Gorji M, Sohrabi M, Tavili A, Pourbabae AA. 2018. Effects of biological soil crusts on some physico-chemical characteristics of rangeland soils of Alagol, Turkmen Sahra, NE Iran. *Soil and Tillage Research*, **181**: 152–159.
- Karnieli A, Ben-Dor E, Patkin K, Banin A. 2002. Mapping of several soil properties using DAIS-7915 hyperspectral scanner data – a case study over clayey soils in Israel. *International Journal of Remote Sensing*, **23**(6): 1043–1062.
- Kaushik BD. 1987. *Laboratory methods for blue green algae*. Associated Publishing Company, New Delhi.
- Kidron GJ. 2016. Linking surface and subsurface properties of biocrusted and non-biocrusted habitats of fine-grained fluvial sediments (playas) from the Negev Desert. *Journal of Hydrology and Hydromechanics*, **64**(2): 141–149.
- Kleiner EF, Harper KT. 1977. Soil properties in relation to cryptogamic ground cover in Canyonlands National Park. *J. Range Manage.*, **30**: 203–205.
- Komárek J. 2005. The modern classification cyanoprokaryotes. *Oceanological and Hydrobiological studies*, XXXIV **3**: 5–17.
- Krumbein WE, Dyer BD. 1985. This Planet is Alive — Weathering and Biology, A Multi-Facetted Problem — . In: *The chemistry of weathering*, Springer, Dordrecht, 143–500.
- Kumar D, Kaštánek P, Adhikary SP. 2016. Diversity of cyanobacteria in biological crusts on arid soils in the Eastern region of India and their molecular phylogeny. *Current Science*, (00113891), **110**(10).
- Lamas BP, Brea MR, Hermo BS. 1995. Colonization by lichens of granite churches in Galicia (northwest Spain). *Science of the Total Environment*, **167**(1–3): 343–351.
- Lane RW, Menon M, McQuaid JB, Adams DG, Thomas AD, Hoon SR, Dougill AJ. 2013. Laboratory analysis of the effects of elevated atmospheric carbon dioxide on respiration in biological soil crusts. *J. Arid Environ.*, **98**: 52–59.
- Lange OL, Meyer A, Zellner H, Heber U. 1994. Photosynthesis and water relations of lichen soil crusts: field measurements in the coastal fog zone of the Namib Desert. *Funct Ecol.*, **8**: 253–264.
- Liu Y, Li X, Xing Z, Zhao X, Pan Y. 2013. Responses of soil microbial biomass and community composition to biological soil crusts in the re-vegetated areas of the Tengger Desert. *Applied Soil Ecology*, **65**: 52–59.
- Luton PE, Wayne JM, Sharp RJ, Riley PW. 1999. The mcrA gene as an alternative to 16S rRNA in the phylogenetic analysis of methanogen populations in landfill. *Microbiology*, **148**: 3521–3530.
- Mager DM, Thomas AD. 2011. Extracellular polysaccharides from cyanobacterial soil crusts: a review of their role in dryland soil processes. *Journal of Arid Environments*, **75**: 91–97.
- Maier S, Muggia L, Kuske CR, Grube M. 2016. *Bacteria and non-lichenized fungi within biological soil crusts biological soil crusts: an organizing principle in drylands*, Springer, 81–100.
- Maqubela M, Mnkeni P, Issa OM, Pardo M, D'Acqui L. 2009. Nostoc cyanobacterial inoculation in South African agricultural soils enhances soil structure, fertility, and maize growth. *Plant and Soil*, **315**(1–2): 79–92.
- McCune B. 1997. *Key to the lichen genera of the Pacific Northwest*. éditeur non identifié.
- Meier LP, Kahr G. 1999. Determination of the cation exchange capacity (CEC) of clay minerals using the complexes of copper (II) ion with triethylenetetramine and tetraethylenepentamine. *Clays and Clay Minerals*, **47**: 386–388.
- Miller MB, Cooper TH, Rust RH. 2011. Diffraction of an eluvial fragipan from dense glacial till in northern Minnesota. *Soil. Sci. Soc. Am. Journ.*, **57**: 787–796.

- Mingorance MD, Barahona E, Fernández-Gálvez J. 2007. Guidelines for improving organic carbon recovery by the wet oxidation method. *Chemosphere*, **68**: 409–413.
- Miralles I, Trasar-Cepeda C, Leirós MC, Gil-Sotres F. 2012. Labile carbon in biological soil crusts in the Tabernas desert, SE Spain. *Soil Biology and Biochemistry*, **58**: 1–8.
- Moniri MH, Sipman HJ. 2009. Lichens of two nature reserves in NE Iran. *Willdenowia*, 199–202.
- Nelson DW, Sommers LE. 1982. Total carbon, organic carbon and organic matter. In: Page AL, Miller RH, Keeney DR (eds.): *Methods of Soil Analysis Part 2, Chemical and Microbiological Properties, second edition*, American Society of Agronomy, Soil Science Society of America: Madison, WI; 539–577.
- Phillips SL, Belnap J. 1998. Shifting carbon dynamics due to the effects of *Bromus tectorum* invasion on biological soil crusts. *Ecol Bull*, **79**, p. 205.
- Prasse R, Bornkamm R. 2000. Effect of microbiotic soil surface crusts on emergence of vascular plants. *Plant Ecology*, **150**: 65–75.
- Qin S, Ferrara KW. 2007. The natural frequency of nonlinear oscillation of ultrasound contrast agents in microvessels. *Ultrasound in medicine & biology*, **33**(7): 1140–1148.
- Rayment GE, Lyons DJ. 2011. *Soil chemical methods: Australasia*, 3. Collingwood: CSIRO.
- Read CG, Popczun EJ, Roske CW, Lewis NS, Schaak RE. 2014. Highly active electrocatalysis of the hydrogen evolution reaction by cobalt phosphide nanoparticles. *Angewandte Chemie*, **126**(21): 5531–5534.
- Rogers SL, Burns RG. 1994. Changes in aggregate stability, nutrient status indigenous microbial populations, and seedling emergence, following inoculation of soil with *Nostoc muscorum*. *Biology and Fertility of Soils*, **18**: 209–215.
- Rozenstein O, Qin Z, Derimian Y, Karnieli A. 2014. Derivation of land surface temperature for Landsat-8 TIRS using a split window algorithm. *Sensors*, **14**(4): 5768–5780.
- Souza-Egipsy V, Wierzechos J, Sancho C, Belmonte A, Ascaso C. 2004. Role of biological soil crust cover in bio-weathering and protection of sandstones in a semi-arid landscape (Torrollones de Gabarda, Huesca, Spain). *Earth Surf. Process. Landforms*, **29**: 1651–1661.
- Thomas AD, Hoon SR, Linton PE. 2008. Carbon dioxide fluxes from cyanobacteria crusted soils in the Kalahari. *Appl. Soil Ecol.*, **39**(3): 254–263.
- Vijayan D, Ray JG. 2015. Ecology and diversity of Cyanobacteria in Kuttanadu paddy wetlands, Kerala, India. *American Journal of Plant Sciences*, **6**(18), p. 2924.
- West NE. 1990. Structure and Function of Soil Microphytic Crusts in Wildland Ecosystems of Arid and Semi-Arid Regions. *Advances in Ecological Research*, **20**: 179–223.
- Wierzechos J. et al. 2012. Microorganisms in desert rocks: the edge of life on Earth. *International Microbiology*, **15**(4): 173–183. doi: 573 10.2436/20.1501.01.170.
- Xu Y, Rossi F, Colica G, Deng S, De Philippis R, Chen L. 2013. Use of cyanobacterial polysaccharides to promote shrub performances in desert soils: a potential approach for the restoration of desertified areas. *Biology and fertility of soils*, **49**(2): 143–152.
- Zaady E, Kuhn U, Wilske B, Sandoval-Soto L, Kesselmeier J. 2000. Patterns of CO<sub>2</sub> exchange in biological soil crusts of successional age. *Soil Biol. Biochem.*, **32**: 959–966.
- Zedda L, Gröngröft A, Schultz M, Petersen A, Mills A, Rambold G. 2011. Distribution patterns of soil lichens across the principal biomes of southern Africa. *Journal of Arid Environments*, **75**(2): 215–220.
- Zhang Y, Aradottir AL, Serpe M, Boeken B. 2016. *Interactions of biological soil crusts with vascular plants. Biological Soil Crusts: An Organizing Principle in Drylands*. Springer, 385–406.
- Zhao L, Yin LJ, Zhai CZ, Chen M, Wang QY, Li LC, Xu ZS, Ma YZ. 2013. Two wheat glutathione peroxidase genes whose products are located in chloroplasts improve salt and H<sub>2</sub>O<sub>2</sub> tolerances in Arabidopsis. *PloS One*, **8**(10): 73989.



---

# Analysis of the difference in depths and variation in slope steepness of the Sunda Trench, Indonesia, east Indian Ocean

**Polina LEMENKOVA<sup>1\*</sup>**

<sup>1</sup>Schmidt Institute of Physics of the Earth, Russian Academy of Sciences, Department of Natural Disasters, Anthropogenic Hazards and Seismicity of the Earth, Laboratory of Regional Geophysics and Natural Disasters (Nr. 303). Bolshaya Gruzinskaya Str. 10, Bld. 1, 123995, Moscow, Russian Federation

Received 19 July 2020; Revised 19 November 2020; Accepted 23 November 2020

\*Correspondence to: Polina LEMENKOVA, e-mail: [pauline.lemenkova@gmail.com](mailto:pauline.lemenkova@gmail.com)

---

## ABSTRACT

The presented paper aims to assess the geomorphology of the Sunda Trench using the scripting cartographic techniques of Generic Mapping Tools (GMT). The GMT presents one of the technical mapping alternatives to the traditional GIS with well-defined scripting approaches for geomorphological mapping and scripting functionality which enables automatization in the geomorphological studies. In this context, modelling of the geospatial datasets is processing using machine learning approaches of GMT which presents fast and accurate modelling of the unreachable regions of the Earth, the deep-sea trenches. In this case, the study focused on the analysis of the submarine geomorphology of the selected segments of the Sunda Trench, located in the Indonesian archipelago, east Indian Ocean, by GMT scripting environment. This research effort is situated in the context of the geomorphology of the Sunda Trench. The aim is to point out which differences can be identified in the slope steepness of the cross-sectional profiles in the selected geomorphological segments of the trench in its southern and northern parts, as distinct by their geological settings and geospatial location. The actuality of the present paper is the analysis of the variations of the geomorphology in Indonesia supported by the GMT-based automated, machine-based geomorphological mapping, which plays an essential role in understanding of the Earth surface processes, supports managements of natural resources, provides information for monitoring and prognosis of natural hazards and enables to predict possible changes in geomorphological evolution of the landforms.

**KEYWORDS** submarine geomorphology; GMT; Sunda Trench; Indian Ocean; geology; cartography; mapping

---

## 1. Introduction

Various methods of geomorphological analysis have been discussed and described so far. To mention a

few of them, these include geodetic measurements (Tregoning et al., 1994), aerial photo-interpretation and terrain analysis (Van Zuidam, 1985) GIS-based applications (Gauger et al., 2007; Bishop et al., 2012;

Pubellier et al., 2003), modelling slopes and trends of the trenches (Lemenkova, 2019f, g). By far the most effective use of cartographic approaches for geomorphological studies is for the cross-section profiling as demonstrating 2D transects of the study area and depicting various geomorphic landforms and slope steepness.

Geomorphological mapping of the submarine features and seafloor is developing rapidly along with cartographic instruments and tools, and computer-based machine learning methods enabling effective data visualization, modelling and mapping. However, besides technical advances, marine geomorphology largely depends on using key progress in general geosciences (geology, environment, landscape studies, tectonics) deriving new data and information from different frontier disciplines of Earth sciences. Such a multidisciplinary approach is highly advantageous for our better understanding of various geomorphological processes recorded at the seafloor. It also results in using different approaches to the geomorphological assessment and various mapping criteria.

A novelty and major asset of the presented manuscript, which is being implemented using Generic Mapping Tools (GMT), is that it presents visualization of the submarine parts of the trench, provides methods of geomorphological modelling and applies quantitative measurements of the depths computations, which not only presents new data on the Indonesian archipelago, but also gives technical approaches that can be repeated in similar geomorphological studies, which has a significant asset due to the automatization of the used cartographic procedures. Hence, the originality of this paper consists in the presented models of the Sunda Trench using GMT, which has not been presented before in the existing literature. Moreover, the new visualized data on the submarine slopes of the Sunda Trench segments can be assimilated in other studies on Indonesian Archipelago and east Indian Ocean basin, as well as applied in the models based on other geospatial datasets (e.g., environmental, marine biological and other thematic studies).

A motivation for developing a present GMT-based geomorphological model of the Indonesian region was the need for automatization techniques

in cartographic routines for the application of GMT modules using more sophisticated algorithms comparing to the traditional GIS. The automatization approach by GMT provides an effective means to visualize the seafloor in geomorphological mapping, an important surface between the geological substrate and the ocean mass where a range of phenomena (biochemical, hydrological, oceanological) are controlled by the geomorphological shape of the submarine landforms and geological character of the seafloor. Although maps of seafloor geomorphology are important for a wide range of science branches (geology, oil and gas engineering, fisheries and marine biological mapping), the cartographic techniques of the submarine geomorphological maps remain a challenge due to the remote location of the study object and the specifics of the GMT syntax. Besides, a submarine geomorphological mapping is strongly limited to the high-resolution data, that is a high-resolution GEBCO/SRTM topography grids.

In this context, work on the analysis and geomorphological modelling of bathymetric data in the Indonesian archipelago was stimulated by the availability of GEBCO dataset. In addition to used high-resolution datasets (GEBCO, EGM-2008 etc), the GMT provides a cartographic methods of the geomorphological mapping through the significant level of the machine-based modelling and advanced solution for graphical visualization of the datasets. In the following parts of the paper a summary of the regional geological settings of the Indonesian archipelago is given, followed by a description of the GMT modelling of the topographic data, the algorithms used for cartographic visualization are discussed by giving the references to the methodology, and geospatial data retrieval through the available open source is described.

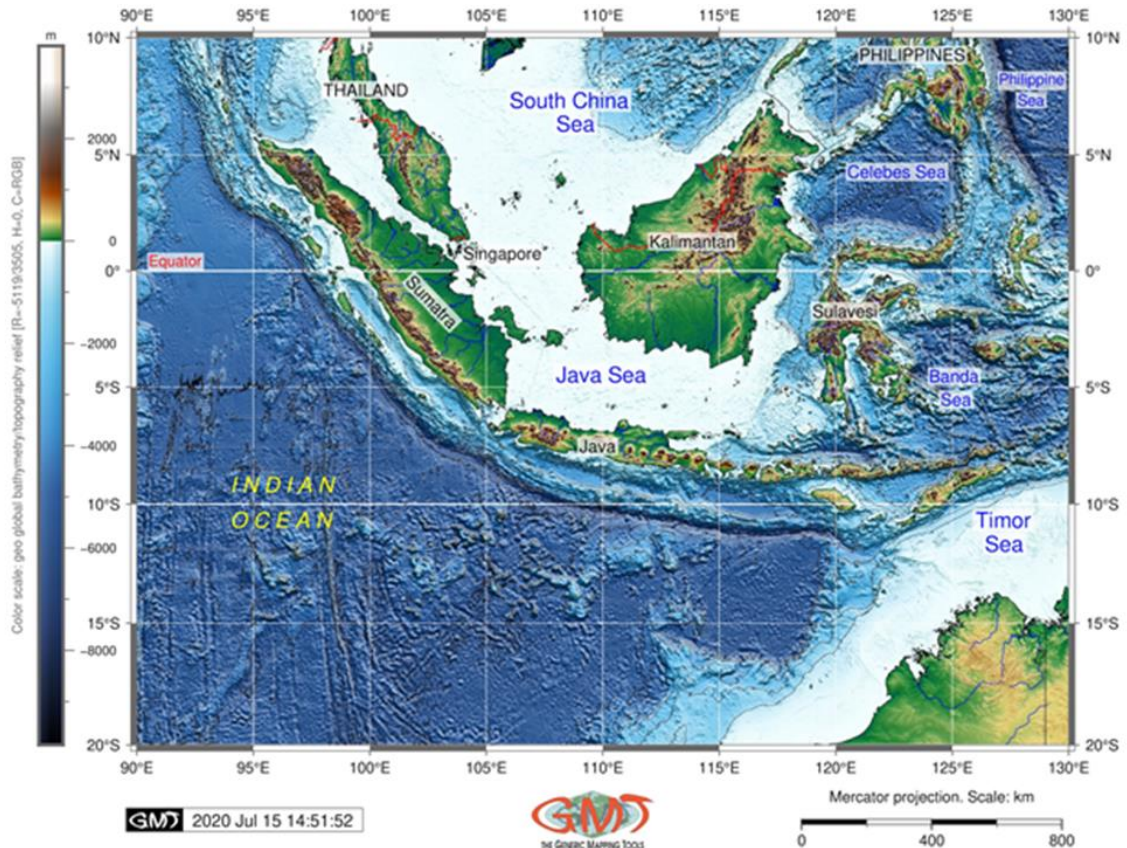
## 2. Study area

### 2.1 Geographic location

The study is focused on the Sunda Trench (also called Java Trench in the past), the deepest oceanic depression of the Indian Ocean, stretching 4,000–5,000 km parallel to the Sunda Island Arc, along its foot. Geographically, it stretches roughly in the NW–

SE direction starting from the Andaman Islands, along the Indonesian archipelago, Sumatra, Java, and Lesser Sunda Islands until the Island of Timor. Its seafloor bottom has the irregular character being wider (up to 50 km) in its northern part and gradually becoming narrow in the southern one (up to 10

km). In particular, the paper analyses the difference in depths and variation in the slope steepness between the two segments of the trench: the southern Java transect (coordinates 108.8°E 10.10°S to 113.0°E 10.75°S) and the northern Sumatra transect (97.5°E 1.1°S to 101.0°E 5.5°S).



**Figure 1** Topographic map of the West Indonesia region basin. GEBCO 15 arc sec resolution global terrain model grid

## 2.2 Bathymetry

The region of the Sunda Volcanic Arc includes the islands of Sumatra and Java, the Sunda Strait between them and the Lesser Sunda Islands stretching between 5 and 10°S (Rahardjo et al., 1995), Fig. 1. The depths of the seafloor in the Sunda Trench increase from the northwest (-3,000 m) to the southeast (-6,000 m), and reach its maximum at -7,209 m. Further to the east, the short Timor Trough (-3,310 m) continues the Sunda Trench marking the boundary between the Indo-Australian Plate and the Timor Plate (Katili, 1972). It is separated by a threshold from the Seram Trough at Kai Islands Arch (-3680 m). The Sunda Island Arc has a complex topographic structure. The northwestern part of the arc

is bordered by the Andaman Sea which has a wide shelf in the east (with depths less than 100 m) and an oval basin in the west. The seafloor bottom of the basin is dissected into several local troughs by a series of the submeridional submarine uphills with depths exceeding -4000 m in its western part.

## 2.3 Geomorphology

The geomorphology framework of the Sunda Trench is largely controlled by the subduction of the Australian plate underneath the Sunda micro-plate. The geological processes take place in basin of the Indian Ocean at different stages of its evolution and influence the nature of the submarine geomorphol-

ogy and geometric shape of the trench. The geomorphology of the Sunda Trench involves the partitioning of the submarine relief of the terrain into conceptual spatial entities based upon morphological criteria, geological processes, rock composition and structure, as well as tectonic evolution of the plates reflected in surface features of the deep-sea trench, and topological relationships of landforms. The submarine geomorphology of the Sunda Trench presents its various landforms and processes in the deep-sea region of the Indonesian archipelago. A variety of local landforms within the trench reflect the action of the selected impact factors at various spatio-temporal scales. These include such factors as tectonic, geological and oceanological ones. The study of the submarine landforms and processes, which largely reflect the evolution of submarine landscapes, has both theoretical and practical values. For example, the seafloor around the oceanic trench includes accumulated renewable and non-renewable resources (e.g. hydrocarbons, deep-sea minerals), which is a practical aim in geological studies. The environmental purpose of the submarine geomorphology consists in the study of habitats of marine species which is necessary for modeling ecosystems among others.

## 2.4 Tectonics

The Java–Sumatra region is notable for the complexity of the tectonic processes in the active convergent tectonic plate margin marked by the subduction of the Indo-Australian Plate under the Eurasian Plate. The complexity of the tectonic setting in the region is illustrated by the presence of three major plates: the Eurasian, the Indian and the Australian with the Australian Plate diving under the Sunda Plate forming a subduction zone. The Indonesian segment of the Eurasian plate has been divided into three minor- and micro-plates that include the following ones: the Sunda Plate, the Banda Sea Plate, the Molucca Plate and the Timor Plate. More details about the tectonic evolution on the region are discussed in the available publications (Barber, 1981; Barber and Crow, 2009; Hamilton, 1977, 1978, 1979; Grevemeyer and Tiwari, 2006). The movement of the Australian plate northwest relative to the Wadati–Benioff zone resulted in the for-

mation of the Sunda Trench–arc system. Local basin depressions of the Indian Ocean formed in the eastern part of the Sunda Island Arc and east of New Guinea where the structures were oriented across the main tectonic plate movement and experienced local rotations. As a consequence, such tectonic processes controlling the evolution of the Indian Ocean give rise to the different geomorphic features as long as they pass through the different periods of the geologic evolution.

The region of Sunda Trench was formed during late Paleozoic (Late Permian) and early Mesozoic (Triassic) periods as a result of the amalgamation of the continental and arc fragments in Indochina, Thailand, Malaysia, and Sumatra (Advokaat et al. 2018). Based on the wide distribution of the Mesozoic marine deposits found on the Sunda Island Arc, there was a deep-sea basin existed between Australia and Indochina during the Jurassic period of Mesozoic. This basin was connected with the Pacific Ocean in the east, and with the Tethys Ocean in the west. The presence of this basin on the territory of modern Sunda Island Arc points at the disintegration of the Gondwana, which occurred at the next period of its geological development. As a result of the seafloor expansion, the oceanic plate subducted under the continental Indo–Australian and Eurasian plates on the eastern border of Tethys during Cretaceous. During Paleogene, the rift zone of Tethys Ocean reduced due to the northward movement of Australia. As a consequence, its seafloor also moved northward, being absorbed in the Wadati–Benioff zone under the system of Sunda Island Arc. The oceanic passage from the Indian Ocean to the Pacific Ocean northward off Australia still existed at the end of the Eocene, although much narrower. Hence, the Sunda Trench–arc system was formed as a result of these complex tectonic movements in course of the geological history of the Indian Ocean region.

## 2.5 Geology

The geological processes and phenomena that take place in the bathymetric basins at different stages of ocean evolution influence the nature of submarine geomorphology and geometric shape of the deep-sea trenches. Tectonic plate subduction and uplift-

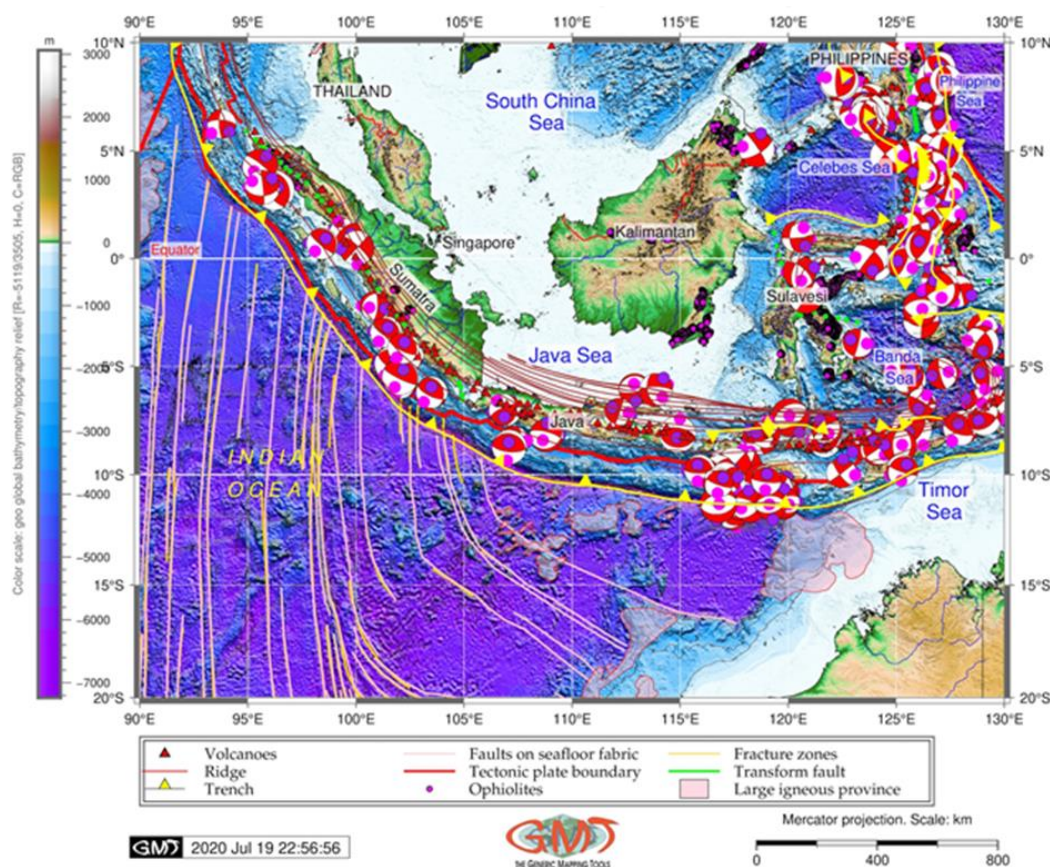
ing driving mechanisms across the Indonesian region have acted remarkably during Cenozoic (Advokaat et al., 2018; Daly et al., 1991; Setijadji et al., 2008; Smyth et al., 2005). The most important, among the many geologic processes that generate and shape the geomorphology of the slopes of the Sunda Trench is plate subduction, which includes bending of the oceanic lithosphere (Karig et al., 1980). Other geological factors affecting the formation of the oceanic trenches include volcanism and sedimentation (Moore et al., 1982; Moore and Curray, 1980; Lemenkova, 2019a, 2018). As a result of the variety of factors, the submarine geomorphology of the Indonesian region was formed. Its geomorphological features include rift valleys, spreading ridges, local minor troughs and a Sunda Trench. The subduction of the plate starts many of kilometers off the trench axis, being caused by the buoyance of the tectonic slab and the elastic nature of the lithosphere (Caméron et al., 1980). The beginning of the plate bending is noted by the outer rise, a bathymetric high on the seaward side of the trench (facing the Indian Ocean, opposite from the Indonesia).

The crystalline basement surface of the Cenozoic structures on the Sunda island arc almost coincides with the modern bathymetry forming the two ridges: the main ridge and the outer ridge, separated by a longitudinal trough with depths of up to 2–4 km. The outer ridge has a steep slope to the Sunda Trench where depths reach up to -6,000 m. The sedimentary cover on the steep slopes of the Sunda Island Arc is generally thin increasing only in local depressions. The thickness of the Cenozoic deposits exceeds 1 km on the seafloor of the Sunda Trench (van Bemmelen, 1970). The lower seafloor layer on the island arcs and submarine ridges, hills and elevations at the bottom of the basin is mainly presented by the granite and metamorphic rocks of the continental crust, especially for the large islands, formed as folded mountain structures. The structure of the rocks of the Sunda Trench–arc system points at Paleozoic schists and gneisses, other metamorphic rocks, granite intrusions and occasional turbidite accumulation in the seafloor of the trench (McDonald, 1977; Anikouchine and Ling, 1967).

## **2.6 Seismicity**

The Sunda Trench is a seismically active part of the Pacific Ring of Fire. A large number of the catastrophic earthquakes are recorded in the Indonesian archipelago, which makes this region especially prone to hazards and risks of the high seismicity (Bellier et al., 1997; Gunawan et al., 2018, 2020; Jena et al., 2020a, b; Marchetti et al., 2020; Moya et al., 2020; NASA, 2018; Socquet et al., 2019; Song et al., 2020; Yulianto et al., 2020). The seismic belt around it connects the Alpide orogenic system and the circum-Pacific seismic belt. A large number of catastrophic earthquakes are recorded and reported in various publications (Abercrombie et al., 2001; Pollitz et al., 2006; Sørensen and Atakan, 2008). The hypocenters of the earthquake surfaces are located under the Sunda Trench and Sunda Island Arc (Nalbant et al., 2005) which ultimately affect the geomorphological shape of the seafloor (Nurwihastuti et al., 2014).

The hypocenters of intermediate earthquakes (focal depth at 60–300 km) and deep-focus earthquakes (focal depth >300 km) form a focal zone with varying inclination: depth of ca. 300 km is 35°, and depth > 300 km is up to 60°, which indicates the fracture of the subducted oceanic plate. The morphology and depth of the subducted plate are defined by the earthquake hypocenters that vary along the plate boundaries (Cardwell and Isacks, 1978). Thus, the crust under the Sunda island arc is up to 25 km thick. It is composed of a thin layer with seismic wave velocities of 3.9–4.7 km/s and much more thick main layers with wave velocities of 5.1–5.7 and 6.6–7.2 km/s typical for solid crustal rocks. Under the seafloor bottom of the Sunda Trench the crustal thickness decreases to 14 km with layers having seismic velocities of 4.1 and 7.2 km/s (Litvin, 1987). More detailed studies on the seismicity and earthquakes in the Indonesian region and Sunda Trench area exist in the published literature (to mention a few, Curray et al., 1977; Harjono et al., 1991; Kieckhefer et al., 1980; Kopp, 2011; Lemenkova, 2020c; Fujii and Satake, 2006; Ammon et al., 2006).



**Figure 2** Geologic map of the West Indonesia basin. GEBCO 15 arc sec resolution global terrain model grid

### 3. Material and Methods

Numerical methods of the data processing are required to model the specifics and reconstruct the past Sumatra and Java tectonic movements. For instance, these include geodetic and paleogeodetic measurements (Chlieh et al., 2008), using GPS and InSAR measurements for estimating land subsidence (Abidin et al., 2008), resolution enhancement techniques for measuring discontinuity depth and correlation with the morphology of the subducting Indo-Australian slab (Dokht et al., 2018). Other approaches include, for example, reconstructed wave heights and ComMIT tsunami modelling (Meservy et al., 2020), method of the radiometric age determination of granitic rocks from the islands for analysis of the geological evolution and geochronology of west Indonesia based on the plate-tectonics concept (Katili, 1973). Development of cartographic methods is presented in works on marine mapping and visualization (Mammerickx et al., 1976), as well as automatization approaches in the GIS techniques

(Schenke and Lemenkova, 2008; Klinger et al., 2012; Lemenkova, 2019e).

The method used to perform cartographic visualization, modelling and mapping is based on the Generic Mapping Tools (GMT). The GMT is developed by Wessel and Smith (1991) and works as a scripting toolset of modules from a shell console. Since 1991 it has been used in geophysical and topographic mapping and being continuously developed since then (Wessel et al., 2013). The GMT has embedded vector shorelines and contours of major geographical objects (rivers, lakes, borderlands) for the World as a basis map used in this research (Wessel and Smith, 1996). For example, a group of modules (such as gmtset, gmtdefaults, grdcut, makecpt, grdimage, psscale, grdcontour, psbasemap, gmtlogo, psconvert and many others) was taken together to perform topographic mapping by a group of sub-tasks: subset raster grid from a raw file, to define projection, visualize, add cartographic elements, and convert the layout out-

put to a graphical format. All this is done sequentially, in a GMT shell script consisting from a sequential line of codes, similar to the principle of programming.

Technical description of the procedure is as follows. First, a subset of the ETOPO1 raster grid has been extracted from the global GEBCO grid using the following code: `"grdcut GEBCO_2019.nc -R105/115/-14/-5.5 -Gstj_relief.nc"`. Then, the data have been checked up (extreme and range in meters) using GDAL library using the following code: `"gdalinfo stj_relief.nc -stats"`. The data for the Java segment demonstrated the following data extent: Minimum=-7239 m, Maximum=3063 m. The data for the Sumatra segment demonstrated following data extent: Minimum=-7164 m, Maximum=3403 m. The next step included making a color palette using the following code: `"gmt makecpt -Cglobe -V -T-7239/3063 > myocean.cpt"`. Afterwards, the file was generated: "ps=crossSTJ.ps" and the raster image was visualized in the given geographic extent here example for the Java segment): `"gmt grdimage stj_relief.nc -Cmyocean.cpt -R105/115/-14/-5.5 -JM6i -P -I+a15+ne0.75 -Xc -K > $ps"`.

The next two steps included adding a color scale by the following code: `"gmt psscale -Dg103.7/-14+w13c/0.4c+v+o0.3/0i+ml -Rstj_relief.nc -J -Cmyocean.cpt -Baf+I"Colors for global bathymetry/topography relief [R=-7164/3403, H=0, C=RGB]" -I0.2 -By+lm -O -K >> $ps"`, and adding cartographic elements (grid, annotated coordinate projection and a time stamp) using the following GMT code: `"gmt psbasemap -R -J -Lx13.0c/-0.5i+c50+w250k+I"Mercator projection. Scale (km)" +f -Bpxg2f1a2 -Bpyg2f1a1 -Bsxg1 -Bsyg1 -BwESN -UBL/-5p/-35p -O -K >> $ps"`. The shorelines and plotted title were visualized using the following GMT code: `"gmt grdcontour stj_relief.nc -R -J -C1000 -B+t"Cross-sectional profiles of the Sunda Trench, Java segment. DEM: GEBCO" -W0.1p -O -K >> $ps"`.

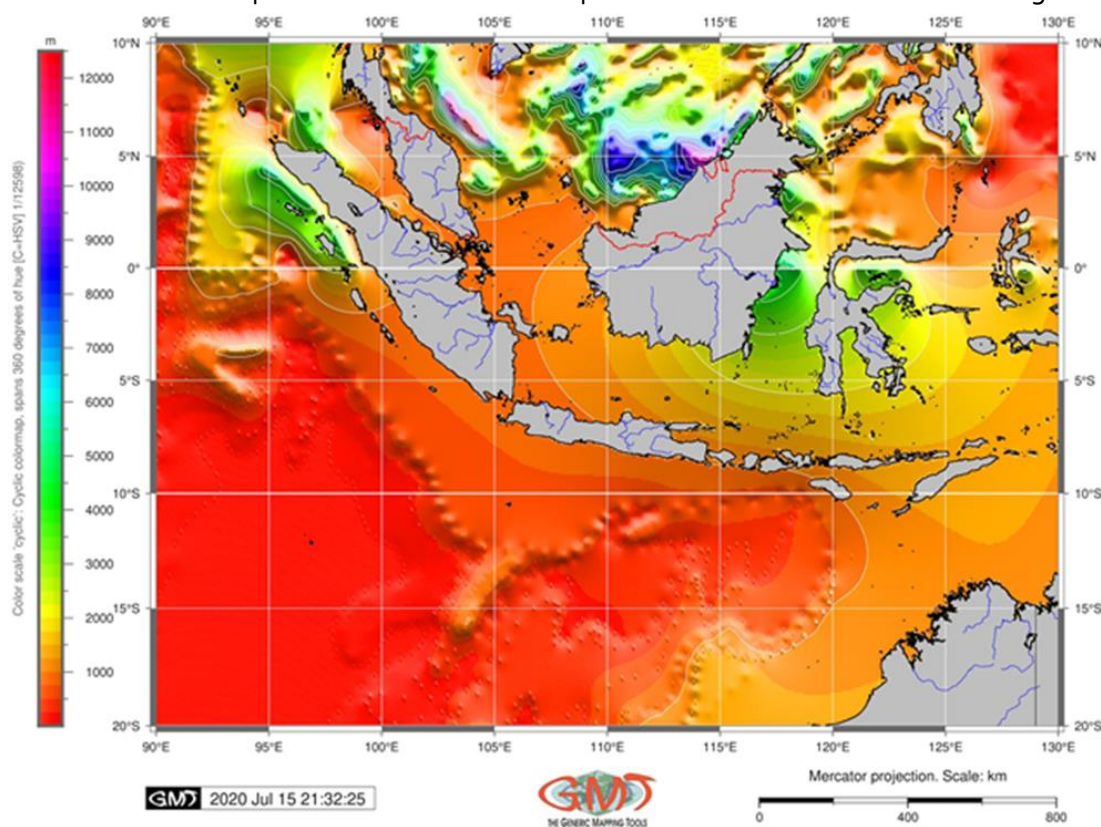
After the cartographic work has been done, the topographic data were used to extract information on depths/heights and model the cross-section profiles. This was done using the following modelling procedure. First, the two points (start point and end point) were selected using UNIX 'cat' utility, as fol-

lows: `cat << EOF > trenchSTJ.txt 108.8 -10.10 113.0 -10.75 EOF` (here EOF signifies "end of file" for the expression). Then the trench segment and end points were plotted (visualized on a map) using the following two chunks of code: 1) `gmt psxy -R -J -W2p,red trenchSTJ.txt -O -K >> $ps` (plotting a line); 2) `gmt psxy -R -J -Sc0.15i -Gred trenchSTJ.txt -O -K >> $ps` (plotting start and end points). Afterwards, a cross-track profiles were generated with the following parameters: 500 km long, sampled every 2km, spaced 10 km and stacked using the mean: `"gmt grdtrack trenchSTJ.txt -Gstj_relief.nc -C500k/2k/10k+v -Sm+sstackSTJ.txt > tableSTJ.txt, gmt psxy -R -J -W0.5p,yellow tableSTJ.txt -O -K >> $ps"`. The data were written in a txt file: `"gmt convert stackSTJ.txt -o0,6 -I -T >> envSTJ.txt"`.

Finally, the graph has been plotted using the following code: Plot graph (statistical median for the profiles): `"gmt psxy -R-250/250/-8500/1000 -JX15.2c/5c -Y15.5c envSTJ.txt -W0.5p -Bpxa50g100f10+I"Distance from trench (km)" -Bpya1000gf500+I"Depth (m)" -Bsxg50 -Bsyg1000 -Glightgray -O -K >> $ps"`. The median lines were added on the graphs using the code: `"gmt psxy -R -J -W1.0p -Ey+p0.2p stackSTJ.txt -O -K >> $ps, gmt psxy -R -J -W1.0p,red stackSTJ.txt -O -K >> $ps"`. Examples of other cartographic scripting are provided in the existing papers, for example GRASS GIS (e.g. Lemenkova, 2020a, 2020b). Comparing to GRASS GIS, a GMT approach is facilitated by a process division into sub-tasks and one or more modules 'responsible' for the executing these tasks that visualize certain cartographic elements (e.g. add annotations, coasts, visualize raster image from a grid, add colour legend, add information about projection, defile grid on the cartographic layout, etc.) and solutions by sketching them out on maps. To a certain extent, this principle can be compared to the structure of layers in a standard GIS menu, e.g. in the ArcGIS (Suetova et al., 2005; Lemenkova et al., 2012; Klaučo et al., 2014, 2017; Lemenkova, 2011). The result of a GMT data processing consists of the print-quality series of maps, geomorphological models, and descriptive statistical analysis. The importance of the precision and accuracy of the raw topographic and geophysical data for mapping has been discussed previously (Smith, 1993; Wessel

and Watts, 1988; Stagpoole et al., 2016; Weatherall et al., 2015). Therefore, high-resolution data were selected as the materials for this research: the topography based on the GEBCO 15-arc-second grid (GEBCO Compilation Group 2020) which is using SRTM basemap (Tozer et al., 2019), geoid based on the 2.5 minute Earth Gravitation Model of 2008 (EGM2008) raster grid (Pavlis et al., 2012), vector layers from the repository of Scripps Institution of Oceanography (SIO), U.S., marine free-air gravity grid (Sandwell et al., 2014; Smith and Sandwell, 1995), data on tectonic plate boundaries and

movements, geological data on earthquakes from the global CMT project (Ekström et al., 2012; Dziewonski et al., 1981). Sediment thickness data were taken from the GlobSed global 5-arc-minute total sediment thickness dataset (Straume et al., 2019). The information on the data layers, extend on topographic range (depths) and projections were retrieved using GDAL (GDAL/OGR contributors, 2020). The cross-section profiling was done using 'grdtrack' module of GMT by automated digitizing of the profile transects crossing the trench in a perpendicular direction in two selected segments.



**Figure 3** Sediment thickness of the West Indonesia basin. GlobSed 5 arc min grid V-3

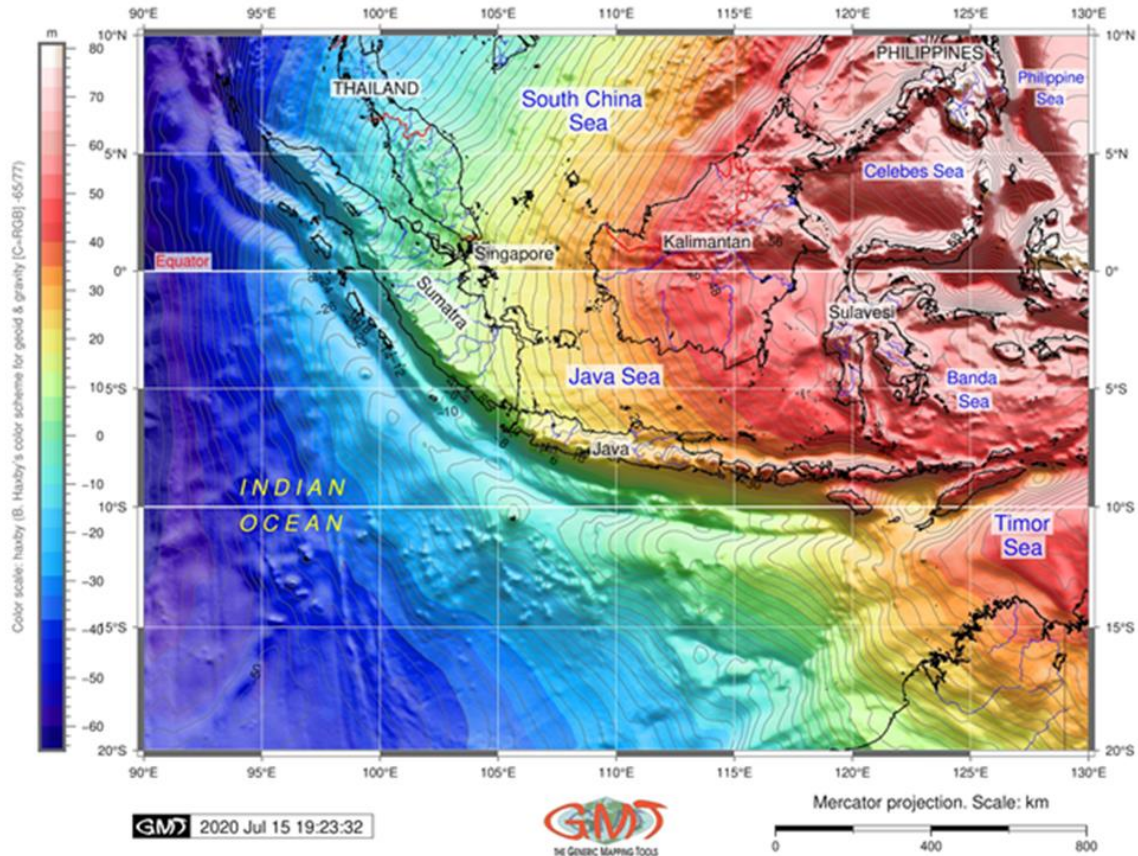
#### 4. Results

As a result of complex geological settings and tectonic evolution, the geomorphology of the Sunda Trench region is varying in its different segments. The outer slope of the Sunda Island Arc has a typical convex profile and stepped structure. The ridge of the Sunda Island Arc is formed here by massive submarine narrow elevations with stepped slopes. In general, the Sunda Island Arc is formed as a double arc stretching in a southeast direction to the New

Guinea. The islands of Sumatra, Java and Lesser Sunda Islands (Sumbawa and Flores and others) are located on its inner ridge. The outer ridge is much narrower with the small islands of Nias Island, located off the western coast of Sumatra, and a chain of Mentawai Islands Regency rising on its western part. The islands of Sumba, Timor, a group of Tanimbar Islands and Seram are located in the east of the Sunda Island Arc (IHO-IOC, 2012). The submarine ridges are separated by a longitudinal trough consisting of the successively located narrow

depressions, with depths gradually increasing from the NW (-1000 to -1500m) to the SE (-3700 to -4000 m). Some more detailed descriptions of the geo-

morphology of the Sunda region supporting this study is given by Verstappen (2000), Curray et al. (1982) and Karig et al. (1979).

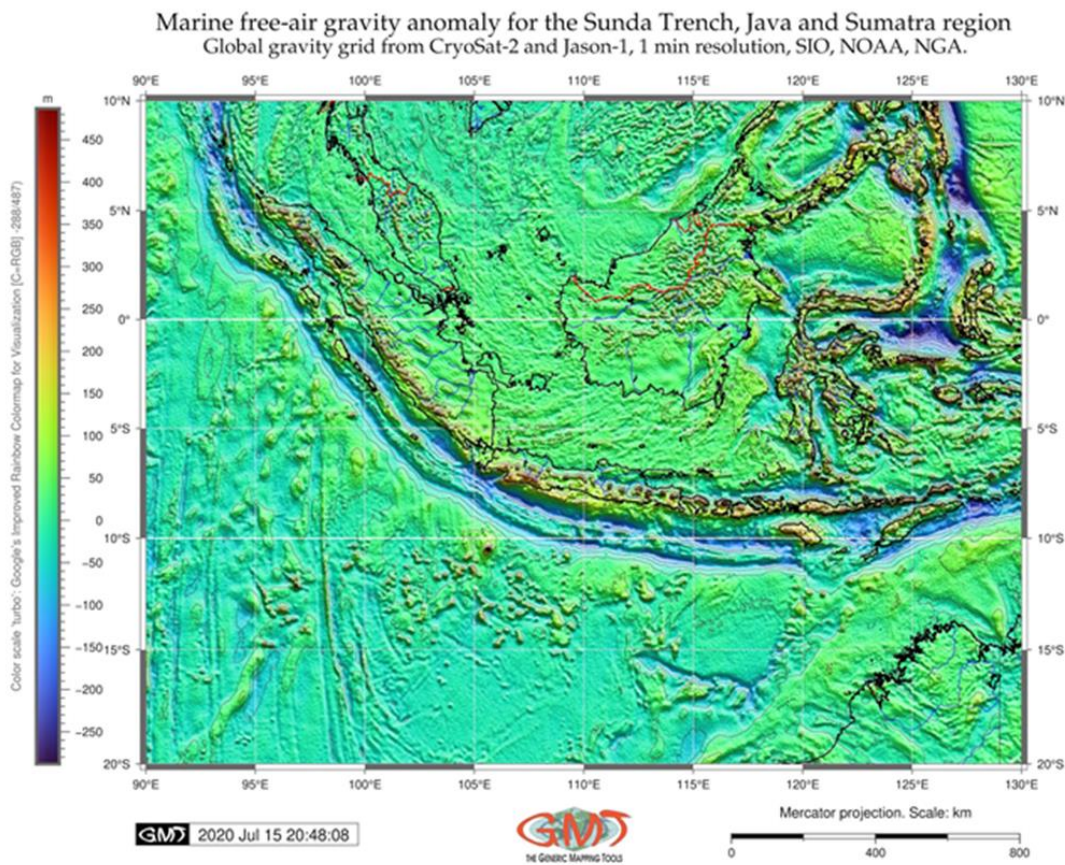


**Figure 4** Geoid model of the West Indonesia basin. World geoid image EGM2008 vertical datum 2.5 min resolution

The seafloor bottom of the trench varies, being different in the segments off Sumatra and Java. In the southeast of Java, it is presented by a series of local depressions separated by the uphills. Comparing to other oceanic trenches, such as Tonga, Kermadec, Vityaz, Vanuatu, Kuril–Kamchatka (Lemenkova, 2019b, 2019c, 2020d), the slopes of the Sunda Trench are generally steeper and asymmetric. The asymmetry of the slopes is reflected in differential geometry of its oceanic and landward sides as follows: comparing to the oceanic slope, the landward slope is higher and steeper, more dissected by the canyons and complicated by the steps and ledges. In the Sumatra and northern Java segment, the seafloor bottom is up to 35 km wide, levelled by a layer of terrigenous sediments with a large ad-

mixture of volcanic material, the thickness of which reaches up to 3 km in the north. The Sunda outer ridge stretches up to several hundred meters in heights, divided into two parts by a saddle along the Sunda Trench. Some block structures and separate seamounts rise with a height of 2,000–3,000 m rise on its convex surface.

The main topographic map of the region is based on the GEBCO grid (Fig. 1). The geologic data included several categories of objects related to the geologic settings of the Sunda Trench and Indonesia region: the location of trench and ridges, volcanoes, slabs, ophiolites, tectonic plate boundaries, focal mechanisms, showing the geological complexity of the region (Fig. 2).

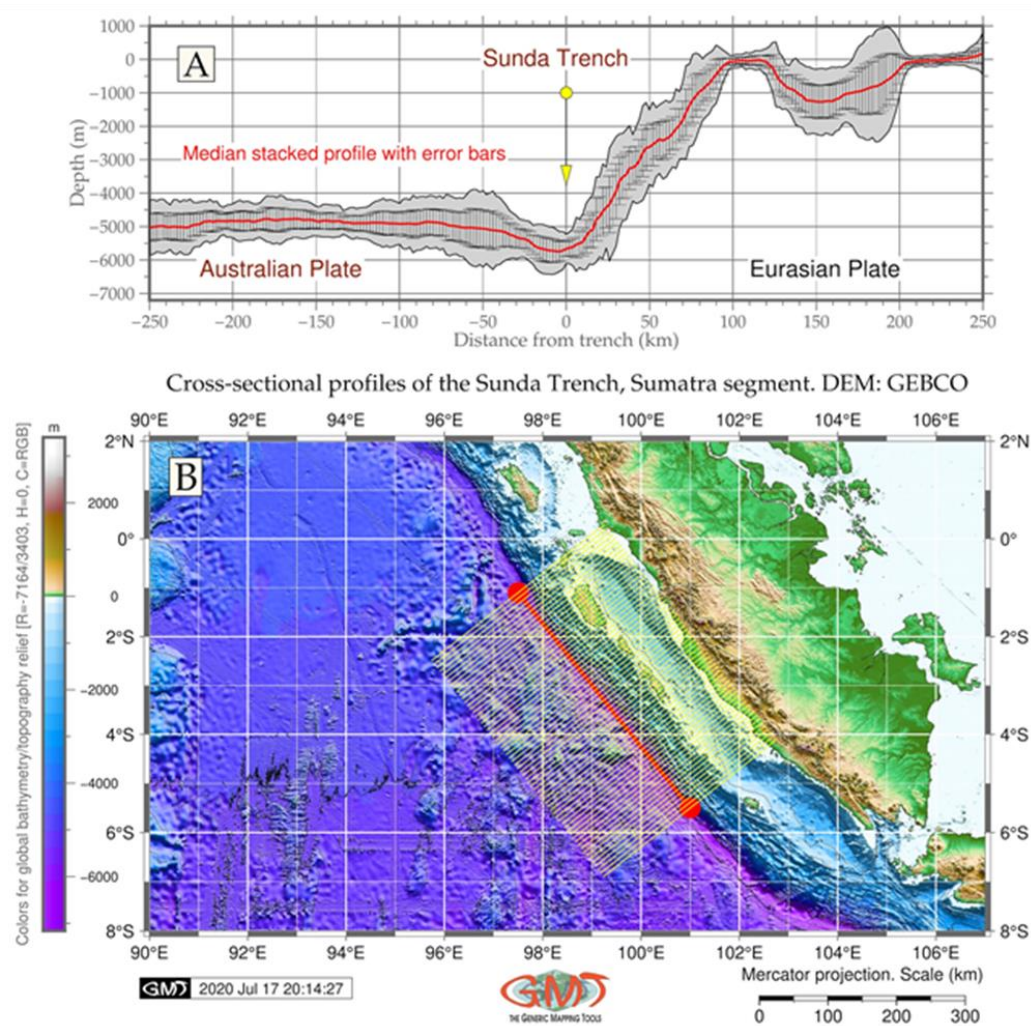


**Figure 5** Marine free-air gravity map of the West Indonesia basin

The sediment thickness map depicts the thickness of the sediment layer with isolines plotted by each 1000 m. The highest sediment thickness can be seen offshore the Kalimantan Island (blue, magenta to purple coloured areas Fig. 3) with values over 8,000 m, with the highest values (magenta colour, >10,000 m) near the Brunei area, Brunei Bay (Fig. 3). The increased values of the sediment thickness can be seen to the south-east off Kalimantan, Sulawesi, north-west Sumatra (green to cyan colours in Fig. 3), values from 3,000 to 7,000 m. That clearly points at the higher values of sediment in the Sumatra segment of the Sunda Trench comparing to the Java segment, where the dominating values are below 2,000 m (orange to red colours in Fig. 3).

The map of the geoid regional model (Fig. 4) clearly shows the asymmetric undulation over the study area: the higher categories of the geoid undulation with values above 45 m (red colours on Fig. 4) are seen in the north-eastern part of the map in

the north-eastern region of the Indonesia Archipelago, the Celebes Sea and the Philippine Sea basin continuing over the terrestrial area of the Kalimantan and the Philippines. Comparing topographic contour (Fig. 1) with the geoid isolines (Fig. 4), one can see the correlation between the geophysical fields and topographic elevations. The area to the west of Kalimantan, Java, Sumatra and Thailand demonstrate a gradual decrease in the geoid values to -10 m (light orange to green colours in Fig. 4). The regions of Sumatra and northern Thailand show negative values decreasing further to the region of the Indian Ocean with values below 45 m (blue to dark blue colours, Fig. 4). As for the region of the Sunda Trench, it shows slightly negative values of the -30 to -10 m across the trench with a slight increase in its Java segment with 0 to -5 m (light aquamarine colour) comparing to Sumatra segment, -5 to -15 m (blueish to light cyan colour), Fig. 4.



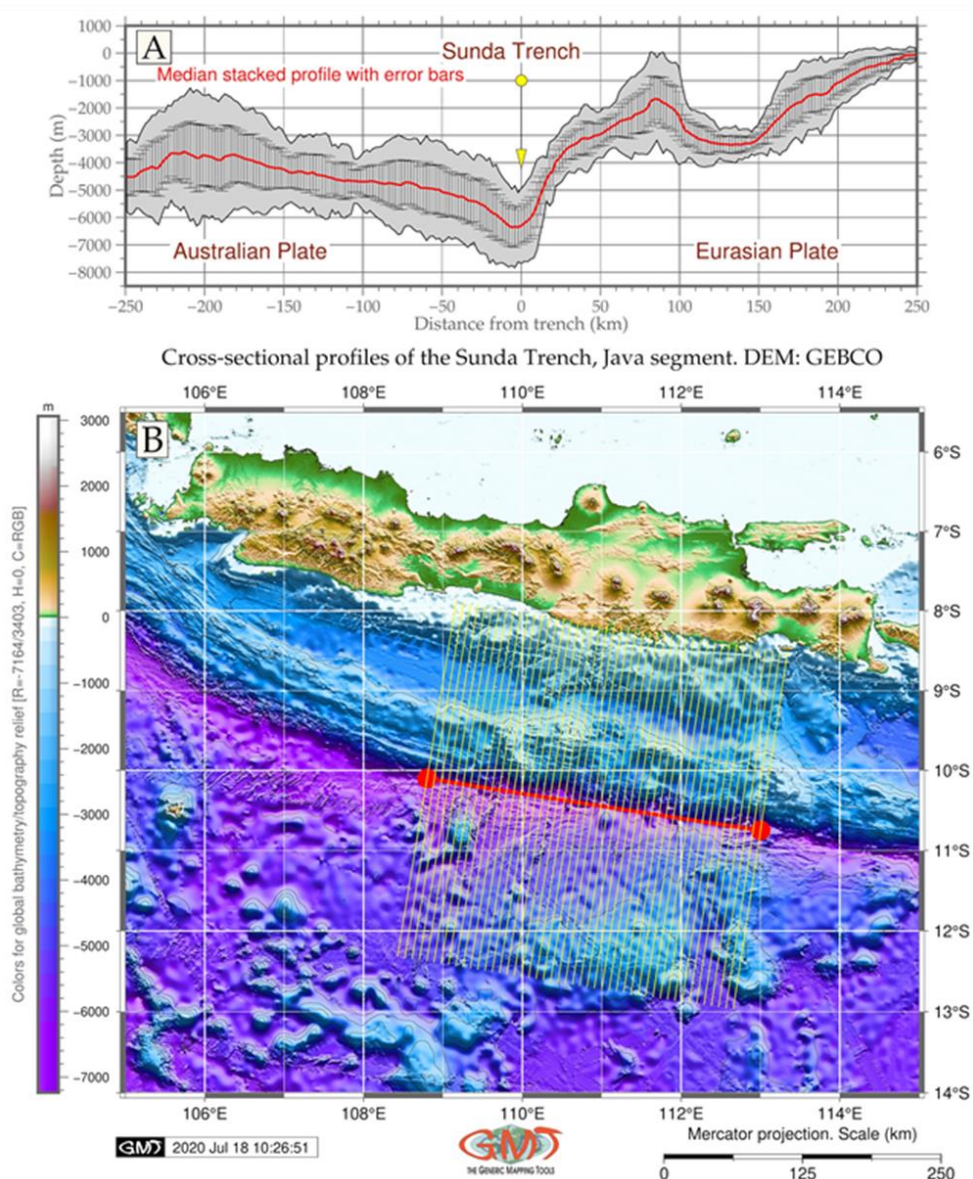
**Figure 6** Cross-section profiles of the Sunda Trench: Sumatra segment

The dominating free-air gravity values are notable for the range of -40 to 40 mGal (light aquamarine colour in Fig. 5). As for the regions of the Sunda Trench, it demonstrates lower values (-40 to -80 mgGal) correlating with the physiographic isolines of the bathymetric map (Fig. 1) which points at the dependences between the distribution of the elevations and the geophysical anomaly fields. The highest values of the marine free-air gravity with >200 mGal (yellow to orange colours in Fig. 5) can be seen over the mountainous areas on Kalimantan and Indonesian archipelago, Sulawesi and central Thailand. The majority of the terrestrial area of Java and Sumatra indicate a correlation between gravity fields and topographic patterns of the mountains areas, Fig. 5.

The selected segments of the Sumatra and Java of the Sunda Trench are selected, since they are spatially distinguishable and formed under the impact of different local geological settings, which introduce a bias in the geomorphological shape of the Sunda Trench and thereby present various parts as more distinct segments. If two closely located segments were selected, the difference in the structure of the trench would have been eliminated and statistically less significant. Therefore, the two segments were selected as the most representative ones. Besides, the Sumatra segment extends in a clear north-west direction, while the Java segment stretches in the north-north-east direction. The comparison of these two segments of the Sunda Trench gives the following results: the Sumatra, which is the northern (Fig. 6) and the Java, which is

the southern (Fig. 7) shows that Java segment has a more symmetrical shape form while the Sumatra segment has a clear asymmetric one-sided shape. The Sumatra segment of the Sunda Trench has a steepness of  $57.86^\circ$  on its eastern side (facing Sumatra Island) and a contrasting  $14.58^\circ$  on the western part facing the Indian Ocean (Fig. 6A). The digitized segment of the Sumatra transect has coordinates  $97.5^\circ\text{E}$   $1.1^\circ\text{S}$  to  $101.0^\circ\text{E}$   $5.5^\circ\text{S}$  (Fig. 6B). The Java segment of the Sunda Trench has a steepness of  $64.34^\circ$  on its northern side (facing Java Island) and  $24.95^\circ$  on the southern part facing the Indian Ocean (Fig. 7A). The digitized segment of Java transect has

coordinates  $108.8^\circ\text{E}$   $10.10^\circ\text{S}$  to  $113.0^\circ\text{E}$   $10.75^\circ\text{S}$ , (Fig. 7B). In both cases the cross-track profiles are plotted as cross-track profiles 500 km long (250 km on each flank from the trench axis), sampled every 2 km, spaced 10 km between each perpendicular line which can be seen in Fig. 6B and 7B as a set of thin, parallel yellow coloured lines. Despite the certain difference in a slope steepness of both flanks of the Java segment (Fig. 7), it has a more symmetric geometry form of the geomorphological profile comparing to the cross-section of the Sumatra segment (northern part of the Sunda Trench, Fig. 6).



**Figure 7** Cross-section profiles of the Sunda Trench: Java segment

The analysis of the statistical histograms (Fig. 8) shows variation in the depths frequency as data distributed along the segments of the Sumatra and Java segments of the Sunda Trench. The Sunda Trench has a bell-shaped data distribution in contrast to the Sumatra segment which has a bimodal (two-peaked pattern) data distribution. The Java segment (Fig. 8, above) has a pool of data concentrating in the most repetitive depths in a bin with a range -2,500 to -5,200 m. The most repetitive data (above 700 samples) are recorded for the following bins: -3,500 to -3,750 m (816 samples), -3,250 to -3,500 m (811 samples), -4,750 to -5,000 m (791 samples), -2,750 to -3,000 m (743 samples) and -3,000 to -3,250 m (721 samples). It shows that in general, the majority of data are concentrated on the depths at -2,500 to -5,200 m.

The Sumatra segment of the Sunda Trench demonstrates a bimodal pattern of the data distribution. Thus, in contrast with the Java transect, the Sumatra transect has the two peaks corresponding to the two intervals: 1) a classic bell-shaped peak at the depths -4,500 m to -5,500 m with values above 600 samples in each bin; 2) a distinct shelf area with a peak from 0 to -1,750 m. The data at the middle depths (between -1,750 to -4,500 m) have a frequency below 300 observation points. The most frequent bathymetric data for the Sumatra segment of the trench correspond to the bin of -4,750 m to -5,000 m (2,151 samples). Comparing to the Sumatra segment, the Java segment is in general deeper. For instance, if comparing the depths below -6,000 m, there are only 138 samples for the Sumatra segment while 547 samples for Java segment. Similarly, the middle -valued depths (those between -2,000 to -4,000 m) have clearly less samples for the Sumatra segment than for the Java segment (compare both plots in Fig. 8). In context of regional topographic variations, the comparison of the histograms points at the difference in depth distribution for the Sumatra and Java segment of the Sunda Trench: a bimodal character of data distribution for a Sumatra segment Trench and a single-peaked data distribution for a Java segment. Similarly, it shows the difference in depths for the both segments: the southern Java segment is deeper comparing to the Sumatra segment, and the geomorphology of the Su-

matra segment has a clear asymmetric view comparing to the more V-shaped Java segment.

## 5. Discussion

The power of GMT-based cartographic visualization, demonstrated in this paper, is primarily relied upon for geospatial analysis, introducing variations in two distinct segments of the Sunda Trench with respect to their local geologic and geophysical settings which explain the terrain segmentation and different slope steepness in two parts of the trench. The paper furthermore discusses the geomorphology of the Sunda Trench, an oceanic trench located in eastern Indian Ocean along the Sumatra and Java Islands of the Indonesian archipelago. The thematic maps and geomorphological modelling were plotted using Generic Mapping Tools (GMT).

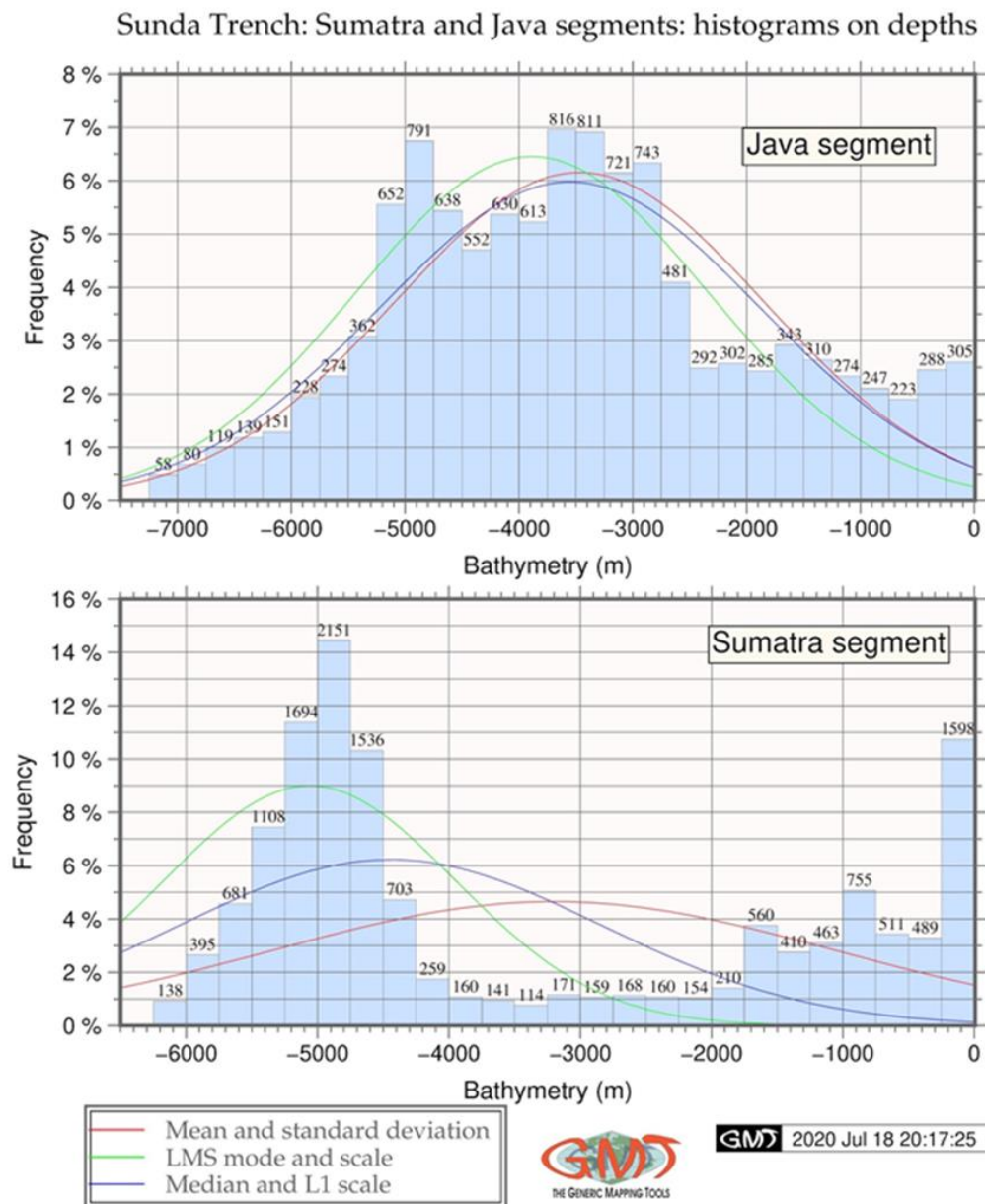
The materials include high-resolution data on topography, geology and geophysics: GEBCO 15 arc-minute resolution grid, EGM2008 2.5 minute Earth Gravitation Model of 2008, GlobSed global 5-arc-minute total sediment thickness and vector geological datasets. In addition to the GEBCO-based bathymetric data, geological, topographic and geophysical maps, the results include enlarged transects for the Java and Sumatra segments, their slope gradients and cross-section profiles, derived from the bathymetric GEBCO dataset.

The presented histograms show variation in depths along the segments of the Sumatra and Java. The Java segment has a bell-shaped data distribution in contrast to the Sumatra with bimodal pattern. The Java segment has the most repetitive depths at -2,500 to -5,200 m. The Sumatra transect has two peaks: 1) a classic bell-shaped peak at depths -4,500 m to -5,500 m; 2) shelf area with a peak from 0 to -1,750 m. The data at middle depths (-1,750 to -4,500 m) have a frequency <300 samples. The most frequent bathymetry for the Sumatra segment corresponds to the -4,750 m to -5,000 m (2,151 samples). Comparing to the Sumatra segment, the Java segment is deeper. For the depths > -6,000 m, there are only 138 samples for the Sumatra while 547 samples for Java. Furthermore, Java segment has more symmetrical geometric shape while Sumatra segment is asymmetric, one-sided. The Sumatra segment has a steepness of 57.86° on its

eastern side (facing Sumatra Island) and a contrasting  $14.58^\circ$  on the western part. The Java segment has a steepness of  $64.34^\circ$  on its northern side (facing Java Island) and  $24.95^\circ$  on the southern part (facing Indian Ocean). The paper contributes to the studies of the submarine geomorphology in Indonesia.

The basin of the Indian Ocean, formed as a result of the long-term evolution is characterized by the presence of the ocean trenches (e.g. Makran Trench, Sunda Trench). As shown in the Fig. 3, the

region has a relatively thick sediment deposits (exceed of 1,000 m), extended areas of continental rise and presence of submarine fans. Although the area of the Sunda Trench is relatively small within the Indian Ocean, it plays a significant role in the geomorphology of its basin (besides other landforms, such as seamounts, plateau of mid-oceanic ridges), serves as a sediment trap and a habitat for the deep-sea fauna, and a unique geomorphic landform.



**Figure 8** Statistical histograms of the of the Sunda Trench

The presented GMT techniques for mapping, cross-section digitizing and spatial modelling are an excellent alternative to classical GIS cartographic methods. In particular, scripting iterative methods enable to produce machine-plotted maps with higher precision compared to that achieved with hand-made methods of manual drawing in GIS. A drawback of a console-based cartographic scripting might be its high learning curve: a GMT does not have a standard GUI such as in ArcGIS (Klaučo et al., 2013a, 2013b) but only a console. However, in case of processing of the large volumes of spatial multi-source data and a need of digitizing the cross-section segments for bathymetric and geomorphological mapping, scripting techniques of GMT represents an excellent solution to reduce the efforts during cartographic routine and to increase the precision of such unreachable areas as the deep-sea trenches.

This paper reviews the new approach of GMT for automated processing of the high-resolution spatial data and geomorphological modelling approaches for plotting cross-section profiles. Although the geomorphological profiles could be done in other GIS software (QGIS, ArcMap etc.), the cartographical layout differs comparing with those approaches, because the GMT-based graphics is made using the machine approach and hence, the cartographic representation of the plots is presented in more details and accuracy comparing to the traditional hand-made digitizing. Through automatization, the GMT permit geomorphologists to go far beyond traditional mapping using GIS software, because it permits a significant degree of machine-learning approaches, a quantitative characterization of the morphology of the submarine landforms and the integration of the thematic datasets (geophysics, geology, sedimentation) for a deeper analysis of the seafloor geomorphology.

## **6. Conclusion**

The paper has given an overview of the presented models of the two segments of the Sunda Trench which contributes to the increase of geomorphological knowledge of this specific region of the Indian Ocean, with emphasis on the Java and Sumatra segments, spatial distinct and differing from each

other. The presented GMT methodology has extended the cartographic application of the scripting methods for geomorphological modelling by including code snippets, and also inserted the examples of the code with a GMT syntax. The interaction of the submarine geomorphology with other settings in the study area (geology, sediments, geophysics) results in their impacts on the actual geomorphological landforms in the submarine part of the Indonesian archipelago and explains slope variations in the northern and southern segments of the trench. The article also discussed in general details the regional settings of the study area that include effects of seismicity and geophysical instability which also affect the landforms in various parts of the trench.

The results from the geomorphological models of the two segments regarding the slope steepness and bathymetric variability agree with assumption that the geologic settings and topography is largely reflected in the geomorphological form of the trench. Apart from the fact that that Java segment has a more symmetrical shape form, while the Sumatra segment has an asymmetric one-sided shape, it was pointed out that their steepness vary accordingly. The effectiveness of the GMT has also been demonstrated: comparing to the GMT, usual traditional GIS software is not effective for modelling of the GEBCO grid due to the large dataset of GEBCO (11,72 Gb for the GEBCO\_2019.nc in a NetCDF format), while the GMT enables to effectively subset the necessary region and process the data subset effectively, precisely and timely. In view of this, the GMT presented an effective cartographic solution for processing big data in geomorphological modelling.

The submarine geomorphology of the oceanic trenches is formed as a result of the complex interactions between the processes of the tectonic plate subduction, geologic setting of the region and geodynamic. As a structural part of the ocean seafloor topography, the evolution of the oceanic trench reflects these changes, which are mirrored in its actual shape (Lemenkova, 2019d). This suggests that submarine topography has mainly a dynamic character, being strongly influenced by both the previous geologic evolution and current external geo-

physical and geodynamic factors. The complexity of the natural phenomena and the connectivity and responses between various geological, geomorphic and climate factors are discussed previously (Moore et al., 1976; Kuhn et al., 2006; Widiyantoro et al., 2011).

The increased availability of high resolution grids (GEBCO, EGM2008, SRTM) enables to better distinguish the genetic origin of the seafloor landforms. Using datasets with a globe coverage allows better understand the context of the submarine geomorphological systems of the oceanic trenches in context of their regional geophysical setting, geological development and tectonic history. Needless to say that this progress has been facilitated by a rapid development of the machine learning techniques, data science approaches and efforts in open data repositories by SIO and GEBCO. Last but not least, the progress in the computer hardware and memory enables to process big datasets for the geomorphological mapping (e.g. original GEBCO grid, 11,72 Gb).

The geomorphological mapping strongly depends on the input data quality and resolution. Thus, the bathymetric data can vary in terms of precision e.g. GEBCO 15 arc-second data acquired for visualizing seafloor versus regional-scale compilations from ETOPO1, ETOPO2 or ETOPO5 with 1, 2 and 5-minute resolution, respectively. Therefore, the quality of the initial bathymetric grids facilitates the delineation of the geomorphological features in the output maps. In particular, it is true for the semi-automated techniques, as demonstrated by the GMT cross-sectional profiling. Higher-resolution topographic raster grids increase the overall quality of the output geomorphological slope profiling. In contrary, poorer resolution cannot ensure the appropriate results in slope modelling. As mentioned before, this research is based on the 15 arc-second resolution topographic GEBCO grid, which ensured the quality and the precision of the results.

The automated data processing approaches of GMT enables time-efficient, precise and accurate cartographic visualization and geomorphological modelling using large data volumes, which now replaces the error-prone traditional, biased, manual methods of data interpretation. Besides, certain

GMT modules ('pshistogram') provide statistical methods of data processing and descriptive statistical analysis. Automated methods of the machine learning provided by GMT, as demonstrated in this paper, significantly reduce mapping subjectivity, labour time and mistakes. Human interpretation can be reduced to the interpretation of the layout outputs and writing the script. The geomorphological interpretation of the landforms can also be used for the additional description of natural complexity and analysis of the correlation with information of the geological and tectonic evolution and geophysical settings of the deep-sea trench. This paper contributed to the regional studies of the Indian Ocean, the geomorphology of the oceanic seafloor and methodological developing of the cartographic approaches in geomorphological mapping.

### Acknowledgements

The author would like to express the gratitude to the two anonymous reviewers for their helpful suggestions, constructive and useful remarks that helped improve a previous version of the typescript. This research has been supported and implemented in the framework of the Project No. 0144-2019-0011, Schmidt Institute of Physics of the Earth, the Russian Academy of Sciences.

### References

- Abercrombie RE, Antolik M, Felzer K, Ekström G. 2001. The 1994 Java tsunami earthquake: Slip over a subducting seamount. *Journal of Geophysical Research*, **106**: 6595–6607. <https://doi.org/10.1029/2000JB900403>
- Abidin HZ, Andreas H, Gamal M, Wirakusumah AD, Darmawan D, Deguchi T, Maruyama Y, 2008. Land subsidence characteristics of the Bandung Basin, Indonesia, as estimated from GPS and InSAR. *Journal of Applied Geodesy*, **2**(3): 167–177. <https://doi.org/10.1515/JAG.2008.019>
- Advokaat EL, Marshall NT, Li S, Spakman W, Krijgsman W, van Hinsbergen DJJ. 2018. Cenozoic rotation history of Borneo and Sundaland, SE Asia revealed by paleomagnetism, seismic tomography, and kinematic reconstruction. *Tectonics*, **37**(8): 2486–2512. <https://doi.org/10.1029/2018TC005010>
- Advokaat EL, Bongers MLM, Rudyawan A, BouDagher-Fadel MK, Langereis CG, Van Hinsbergen DJJ. 2018. Early Cretaceous origin of the Woyla Arc (Sumatra,

- Indonesia) on the Australian plate. *Earth and Planetary Science Letters*, **498**: 348–361.  
<https://doi.org/10.1016/j.epsl.2018.07.001>
- Ammon CJ, Kanamori H, Lay T, Velasco AA. 2006. The 17 July 2006 Java tsunami earthquake. *Geophysical Research Letters*, **33**: L24308.  
<https://doi.org/10.1029/2006GL028005>
- Anikouchine WA, Ling H. 1967. Evidence for turbidite accumulation in trenches in the Indo-Pacific region. *Marine Geology*, **5**: 141–154.
- Barber AJ. 1981. Structural interpretation of the island of Timor, eastern Indonesia. In: E.J. Barber and S. Wirjosujono Eds. The geology and tectonics of eastern Indonesia. *GRDC Special Publications*, **2**: 183–198.
- Barber AJ, Crow MJ. 2009. Structure of Sumatra and its implications for the tectonic assembly of Southeast Asia and the destruction of Paleotethys. *Island Arc*, **18**(1): 3–20. <https://doi.org/10.1111/j.1440-1738.2008.00631.x>
- Bellier O., Sébrier M, Pramumijoyo S, Beaudouin T, Harjono H, Bahar I, Forni O. 1997. Paleoseismicity and seismic hazard along the ornat Sumatran Fault (Indonesia). *Journal of Geodynamics*, **24**: 169–183.  
[https://doi.org/10.1016/S0264-3707\(96\)00051-8](https://doi.org/10.1016/S0264-3707(96)00051-8)
- Van Bemmelen RW. 1970. *The Geology of Indonesia*. 2<sup>nd</sup> edition, Martinus Nijhoff, The Hague.
- Bishop MP, James LA, Shroder Jr JF, Walsh SJ. 2012. Geospatial Technologies and Digital Geomorphological Mapping: Concepts, Issues and Research. *Geomorphology*, **137**: 5–26.  
<https://doi.org/10.1016/j.geomorph.2011.06.027>
- Caméron NR, Clarke MCG, Aldiss DT, Aspden JA, Djunuddin A. 1980. The geological evolution of northern Sumatra. *Proceedings Indonesian Petrology Association, 9<sup>th</sup> Annual Convention*, 149–187.
- Cardwell RK, Isacks BL. 1978. Geometry of the subducted lithosphere beneath the Banda Sea in eastern Indonesia from seismicity and fault plane solutions. *Journal of Geophysical Research*, **83**(B6): 2825–2838.  
<https://doi.org/10.1029/JB083iB06p02825>
- Chlieh M, Avouac JP, Sieh K, Natawidjaja DH, Galetzka J, 2008. Heterogeneous coupling of the Sumatran megathrust constrained by geodetic and paleogeodetic measurements. *Journal of Geophysical Research*, **113**: B05305.  
<https://doi.org/10.1029/2007JB004981>
- Curry JR, Emmel FJ, Moore DG, Raitt RW. 1982. Structure, Tectonics, and Geological History of the Northeastern Indian Ocean. In: Nairn AEM, Stehli FG. (eds) *The Ocean Basins and Margins*. Springer, Boston, MA.
- Curry JR, Shor GG Jr, Raitt RW, Henry M. 1977. Seismic refraction and reflection studies of crustal structure of the Eastern Sunda and Western Banda Arcs. *Journal of Geophysical Research*, **82**: 2479–2489.  
<https://doi.org/10.1029/JB082i017p02479>
- Daly MC, Cooper MA, Wilson I, Smith DG, Hooper BGD. 1991. Cenozoic plate tectonics and basin evolution in Indonesia. *Marine and Petroleum Geology*, **8**(1): 2–21.  
[https://doi.org/10.1016/0264-8172\(91\)90041-X](https://doi.org/10.1016/0264-8172(91)90041-X)
- Dokht, RMH, Gu YJ, Sacchi MD. 2018. Migration imaging of the Java subduction zones. *Journal of Geophysical Research: Solid Earth*, **123**(2): 1540–1558.  
<https://doi.org/10.1002/2017JB014524>
- Dziewonski AM, Chou TA, Woodhouse JH. 1981. Determination of earthquake source parameters from waveform data for studies of global and regional seismicity. *Journal of Geophysical Research*, **86**(B4): 2825–2852. <https://doi.org/10.1029/JB086iB04p02825>
- Ekström G, Nettles M, Dziewoński AM. 2012. The global CMT project 2004–2010: Centroid-moment tensors for 13,017 earthquakes. *Physics of the Earth and Planetary Interiors*, **200**: 1–9.  
<https://doi.org/10.1016/j.pepi.2012.04.002>
- Fujii Y, Satake K, 2006. Source of the July 2006 West Java tsunami estimated from tide gauge records. *Geophysical Research Letters*, **33**: L24317.  
<https://doi.org/10.1029/2006GL028049>
- Gauger S, Kuhn G, Gohl K, Feigl T, Lemenkova P, Hillenbrand C. 2007. Swath-bathymetric mapping. *Reports on Polar and Marine Research*, **557**: 38–45.  
<https://doi.org/10.6084/m9.figshare.7439231>
- GEBCO Compilation Group 2020. GEBCO 2020 Grid. <https://doi.org/10.5285/a29c5465-b138-234d-e053-6c86abc040b9>
- GDAL/OGR contributors 2020. GDAL/OGR Geospatial Data Abstraction software Library. Open Source Geospatial Foundation. <https://gdal.org> [online accessed: 17 July 2020].
- Grevemeyer I, Tiwari VM. 2006. Overriding Plate Controls and Spatial Distribution Of Megathrust Earthquakes in the Sunda–Andaman Subduction Zone. *Earth and Planetary Science Letters*, **251**: 199–208.  
<https://doi.org/10.1016/j.epsl.2006.08.021>
- Gunawan E, Widiyantoro S, Rosalia S, Daryono MR, Meilano I, Supendi P, Ito T, Tabei T, Kimata F, Ohta Y, Ismail N. 2018. Coseismic slip distribution of the 2 July 2013 Mw 6.1 aceh, Indonesia, earthquake and its tectonic implications. *Bulletin of the Seismological Society of America*, **108**(4): 1918–1928,  
<https://doi.org/10.1785/0120180035>
- Gunawan E, Widiyantoro S, Supendi P, Nishimura T. 2020. Identifying the most explainable fault ruptured of the 2018 Palu–Donggala earthquake in Indonesia using coulomb failure stress and geological field report.

- Geodesy and Geodynamics*, **11**: 252–257.  
<https://doi.org/10.1016/j.geog.2020.04.004>
- Hamilton W. 1977. Subduction in the Indonesian Region. In Talwani, M. and Pittman, W. C. (eds.), *Island Arcs, Deep-Sea Trenches and Back-Arc Basins*, *American Geophysical Union Maurice Ewing Series*, **1**: 15–31.
- Hamilton W. 1978. *Tectonic map of the Indonesian region*. U. S. Geological Survey Map I-875-D.
- Hamilton W. 1979. Tectonics of the Indonesian Region. *U.S. Geological Survey Professional Paper*, **1078**: 345.  
<https://doi.org/10.3133/pp1078>
- Harjono H, Diamant M, Dubois J, Larue M, 1991. Seismicity of the Sunda strait: evidence for crustal extension and volcanological implications. *Tectonics*, **10**: 17–30. <https://doi.org/10.1029/90TC00285>
- IHO–IOC. 2012. *GEBCO Gazetteer of Undersea Feature Names*.
- Jena R, Pradhan B, Beydoun G, Alamri AM, Ardiansyah, Nizamuddin, Sofyan H. 2020a. Earthquake hazard and risk assessment using machine learning approaches at Palu, Indonesia. *Science of The Total Environment*, **749**: 141582.  
<https://doi.org/10.1016/j.scitotenv.2020.141582>
- Jena R, Pradhan B, Beydoun G, Nizamuddin, Ardiansyah, Sofyan H, Affan M. 2020b. Integrated model for earthquake risk assessment using neural network and analytic hierarchy process: Aceh province, Indonesia. *Geoscience Frontiers*, **11**: 613–634.  
<https://doi.org/10.1016/j.gsf.2019.07.006>
- Karig DE, Moore GF, Curray JR, Lawrence MB. 1980. Morphology and shallow structure of the trench slope off Nias Island, Sunda Arc. In: Hayes DE. (ed.). *The Tectonic and Geologic Evolution of Southeast Asian Seas and Islands*, *Geophysical Monograph Series* Washington, **23**: 179–208. ISBN: 978-1-118-66379-0, 326 pp.
- Karig DE, Suparka S, Moore GF, Hehanussa PE. 1979. Structure and Cenozoic Evolution of the Sunda Arc in the Central Sumatra Region'. In Watkins, J. S., Montadert, L., and Dickerson, P. W. (eds.), *Geological and Geophysical Investigations of Continental Slopes and Rises*, *American Association of Petroleum Geologists, Memoir*, **29**: 223–237.
- Katili JA. 1972. Plate Tectonics of Indonesia with Special Reference to the Sundaland Area. *Proceedings of the 1st Annual Convention Indonesian Petroleum Association*: 57–61.  
<https://doi.org/10.29118/IPA.1921.57.61>
- Katili JA. 1973. Geochronology of west Indonesia and its implication on plate tectonics. *Tectonophysics*, **19**: 195–212. [https://doi.org/10.1016/0040-1951\(73\)90019-X](https://doi.org/10.1016/0040-1951(73)90019-X)
- Kieckhefer RM, Shor GGJr, Curray JR, Sugiarta W, Hehuwat F. 1980. Seismic Refraction Studies of the Sunda Trench and Forearc Basin. *Journal of Geophysical Research*, **85**: 863–889.  
<https://doi.org/10.1029/JB085iB02p00863>
- Klaučo M, Gregorová B, Stankov U, Marković V, Lemenkova P. 2013a. Determination of ecological significance based on geostatistical assessment: a case study from the Slovak Natura 2000 protected area. *Central European Journal of Geosciences*, **5**(1): 28–42.  
<https://doi.org/10.2478/s13533-012-0120-0>
- Klaučo M, Gregorová B, Stankov U, Marković V, Lemenkova P. 2013b. Interpretation of Landscape Values, Typology and Quality Using Methods of Spatial Metrics for Ecological Planning. *54<sup>th</sup> International Conference Environmental & Climate Technologies*. Riga, Latvia.  
<https://doi.org/10.13140/RG.2.2.23026.96963>
- Klaučo M, Gregorová B, Stankov U, Marković V, Lemenkova P. 2014. Landscape metrics as indicator for ecological significance: assessment of Sitno Natura 2000 sites, Slovakia. *Ecology and Environmental Protection*, 85–90.  
<https://doi.org/10.6084/m9.figshare.7434200>
- Klaučo M, Gregorová B, Stankov U, Marković V, Lemenkova P. 2017. Land planning as a support for sustainable development based on tourism: A case study of Slovak Rural Region. *Environmental Engineering and Management Journal* **2**(16): 449–458.  
<https://doi.org/10.30638/eemj.2017.045>
- Klinger T, Heipke C, Ott N, Schenke HW, Ziemis M. 2012. Automated Extraction of the Antarctic Coastline Using Snakes. A special joint symposium of ISPRS Technical Commission IV & AutoCarto in conjunction with ASPRS/CaGIS 2010 Fall Specialty Conference November 15–19, 2010 Orlando, Florida.
- Kopp H. 2011. The Java convergent margin: structure, seismogenesis and subduction processes, *Geological Society, London, Special Publications* **355**: 1, 111–137.  
<https://doi.org/10.1144/SP355.6>
- Kuhn G, Hass C, Kober M, Petitat M, Feigl T, Hillenbrand CD, Kruger S, Forwick M, Gauger S, Lemenkova P. 2006. The response of quaternary climatic cycles in the South–East Pacific: development of the opal belt and dynamics behavior of the West Antarctic ice sheet. In: Gohl K. (ed.). *Expeditions programme 75 ANT XXIII/4*.  
<https://doi.org/10.13140/RG.2.2.11468.87687>
- Lemenkova P. 2020a. Fractal surfaces of synthetical DEM generated by GRASS GIS module r.surf.fractal from ETOPO1 raster grid. *Journal of Geodesy and Geoinformation*, **7**(1):86–102.  
<https://doi.org/10.9733/JGG.2020R0006.E>

- Lemenkova P. 2020b. NOAA Marine Geophysical Data and a GEBCO Grid for the Topographical Analysis of Japanese Archipelago by Means of GRASS GIS and GDAL Library. *Geomatics and Environmental Engineering*, **14**(4), 25–45.  
<https://doi.org/10.7494/geom.2020.14.4.25>
- Lemenkova P. 2020c. Variations in the bathymetry and bottom morphology of the Izu–Bonin Trench modelled by GMT. *Bulletin of Geography. Physical Geography Series* **18**(1): 41–60.  
<https://doi.org/10.2478/bgeo-2020-0004>
- Lemenkova P. 2020d. GMT Based Comparative Geomorphological Analysis of the Vityaz and Vanuatu Trenches, Fiji Basin. *Geodetski List*, **74**(1): 19–39.  
<https://doi.org/10.6084/m9.figshare.12249773>
- Lemenkova P. 2019a. Statistical Analysis of the Mariana Trench Geomorphology Using R Programming Language. *Geodesy and Cartography*, **45**(2): 57–84.  
<https://doi.org/10.3846/gac.2019.3785>
- Lemenkova P. 2019b. Topographic surface modelling using raster grid datasets by GMT: example of the Kuril–Kamchatka Trench, Pacific Ocean. *Reports on Geodesy and Geoinformatics*, **108**: 9–22.  
<https://doi.org/10.2478/rgg-2019-0008>
- Lemenkova P. 2019c. GMT Based Comparative Analysis and Geomorphological Mapping of the Kermadec and Tonga Trenches, Southwest Pacific Ocean. *Geographia Technica*, **14**(2): 39–48.  
[https://doi.org/10.21163/GT\\_2019.142.04](https://doi.org/10.21163/GT_2019.142.04)
- Lemenkova P. 2019d. Geophysical Modelling of the Middle America Trench using GMT. *Annals of Valahia University of Targoviste. Geographical Series*, **19**(2): 73–94. <https://doi.org/10.6084/m9.figshare.12005148>
- Lemenkova P. 2019e. AWK and GNU Octave Programming Languages Integrated with Generic Mapping Tools for Geomorphological Analysis. *GeoScience Engineering*, **65**(4): 1–22. <https://doi.org/10.35180/gse-2019-0020>
- Lemenkova P. 2019f. Testing Linear Regressions by StatsModel Library of Python for Oceanological Data Interpretation. *Aquatic Sciences and Engineering*, **34**: 51–60. <https://doi.org/10.26650/ASE2019547010>
- Lemenkova P. 2019g. An Empirical Study of R Applications for Data Analysis in Marine Geology. *Marine Science and Technology Bulletin*, **8**(1): 1–9.  
<https://doi.org/10.33714/masteb.486678>
- Lemenkova P. 2018. R scripting libraries for comparative analysis of the correlation methods to identify factors affecting Mariana Trench formation. *Journal of Marine Technology and Environment*, **2**: 35–42.  
<https://doi.org/10.6084/m9.figshare.7434167>
- Lemenkova P, Promper C, Glade T. 2012. Economic Assessment of Landslide Risk for the Waidhofen a.d. Ybbs Region, Alpine Foreland, Lower Austria. In: Eberhardt E, Froese C, Turner AK, Leroueil S. (eds.) Protecting Society through Improved Understanding. 11<sup>th</sup> International Symposium on Landslides & the 2<sup>nd</sup> North American Symposium on Landslides & Engineered Slopes (NASL), June 2–8, 2012. Banff, Canada, 279–285.  
<https://doi.org/10.6084/m9.figshare.7434230>
- Lemenkova P. 2011. *Seagrass Mapping and Monitoring Along the Coasts of Crete, Greece*. M.Sc. Thesis. Netherlands: University of Twente. 158 p.  
<https://doi.org/10.13140/RG.2.2.16945.22881>
- Litvin VM. 1987. *Morphostructure of the ocean seafloor*. Nedra, Leningrad.
- Marchetti D, De Santis A, Shen X, Campuzano SA, Perrone L, Piscini A, Di Giovambattista R, Jin S, Ippolito A, Cianchini G, Cesaroni C, Sabbagh D, Spogli L, Zhima Z, Huang J. 2020. Possible Lithosphere–Atmosphere–Ionosphere Coupling effects prior to the 2018 Mw = 7.5 Indonesia earthquake from seismic, atmospheric and ionospheric data. *Journal of Asian Earth Sciences*, **188**: 104097.  
<https://doi.org/10.1016/j.jseaes.2019.104097>
- Mammerickx J, Fisher RL, Emmel FJ, Smith SM. 1976. Bathymetry of the East and Southeast Asian Seas. *Geological Society of America, Map & Chart Series* MC-17.
- McDonald JM. 1977. *Sediments and Structure of the Nicobar Fan, Northeast Indian Ocean*. Unpub. Ph. D. Thesis, U.C.S.D., 148 p.
- Meservy W, Harris R, Setiadi G, Hapsoro S. 2020. Coastal boulder fields and tsunami hazards of East Java, Indonesia. *EGU General Assembly 2020*, Online, 4–8 May 2020, EGU2020–22118,  
<https://doi.org/10.5194/egusphere-egu2020-22118>
- Moore DG, Curray JR, Emmel FJ. 1976. Large submarine slide (olistostrome) associated with Sunda Arc subduction zone, northeast Indian Ocean. *Marine Geology*, **21**: 211–226. [https://doi.org/10.1016/0025-3227\(76\)90060-8](https://doi.org/10.1016/0025-3227(76)90060-8)
- Moore GF, Curray JR. 1980. Structure of the Sunda Trench lower slope off Sumatra from multichannel seismic reflection data. *Marine Geophysical Research*, **4**: 319–340. <https://doi.org/10.1007/BF00369106>
- Moore GF, Curray JR, Emmel FJ. 1982. Sedimentation in the Sunda Trench and forearc region. *Geological Society, London, Special Publications*, **10**: 245–258.  
<https://doi.org/10.1144/GSL.SP.1982.010.01.16>
- Moya L, Muhari A, Adriano B, Koshimura S, Mas E, Marval-Perez LR, Yokoya N. 2020. Detecting urban changes using phase correlation and ℓ1-based sparse model for early disaster response: A case study of the 2018 Sulawesi Indonesia earthquake–tsunami. *Remote Sensing of Environment*, **242**: 111743.

- <https://doi.org/10.1016/j.rse.2020.111743>
- Nalbant SS, Steacy S, Sieh K, Natawidjaja D, McCloskey J. 2005. Earthquake risk on the Sunda Trench. *Nature*, **435**: 756–757. <https://doi.org/10.1038/nature435756a>
- NASA. (2018). Satellite-derived map of ground deformation from earthquake beneath Lombok, Indonesia. <https://disasters.nasa.gov/lombok-indonesia-earthquake-2018/satellite-derived-map-ground-deformation-earthquake-beneath-lombok> Accessed on November 19, 2020.
- Nurwihastuti DW, Sartohadi J, Mardiatno D, Nehren U, Restu. 2014. Understanding of Earthquake Damage Pattern through Geomorphological Approach: A Case Study of 2006 Earthquake in Bantul, Yogyakarta, Indonesia. *World Journal of Engineering and Technology*, **02**(3B): 61–70. <https://doi.org/10.4236/wjet.2014.23B010>
- Pavlis NK, Holmes SA, Kenyon SC, Factor JK. 2012. The development and evaluation of the Earth Gravitational Model 2008 (EGM2008). *Journal of Geophysical Research*, **117**: B04406, <https://doi.org/10.1029/2011JB008916>
- Pollitz FF, Banerjee P, Bürgmann R, Hashimoto M, Choosakul N. 2006. Stress changes along the Sunda Trench following the 26 December 2004 Sumatra–Andaman and 28 March 2005 Nias earthquakes. *Geophysical Research Letters*, **33**: L06309. <https://doi.org/10.1029/2005GL024558>
- Pubellier M, Ego F, Chamot–Rooke N, Rangin C. 2003. The building of pericratonic mountain ranges: structural and kinematic constraints applied to GIS–based reconstructions of SE Asia. *Bulletin de la Société Géologique France*, **174**(6): 561–584. <https://doi.org/10.2113/174.6.561>
- Rahardjo W, Sukandarrumidi, Rosidi HMD. 1995. *Geological Map of the Yogyakarta Sheet, Jawa, Scale 1:100.000*. Geological Research and Development Centre, Bandung.
- Sandwell DT, Müller RD, Smith WHF, Garcia E, Francis R. 2014. New global marine gravity model from CryoSat–2 and Jason–1 reveals buried tectonic structure. *Science*, **346**(6205): 65–67. <https://doi.org/10.1126/science.1258213>
- Schenke HW, Lemenkova P. 2008. Zur Frage der Meeresboden–Kartographie: Die Nutzung von AutoTrace Digitizer für die Vektorisierung der Bathymetrischen Daten in der Petschora–See. *Hydrographische Nachrichten*, **81**: 16–21. <https://doi.org/10.6084/m9.figshare.7435538>
- Setijadji LD, Watanabe K, Barianto DH, Rahardjo W, Sudarno I, Susilo A, Itaya T. 2008. Searching for the Active Fault of the Yogyakarta Earthquake of 2006 Using Data Integration on Aftershocks, Cenozoic Geo–History, and Tectonic Geomorphology. In: Karnawati D, Pramumijoyo S, Anderson R, Husein S. (eds.) *The Yogyakarta Earthquake of May 27, 2006*. Star Publisher, Los Angeles.
- Smith WHF. 1993. On the accuracy of digital bathymetric data. *Journal of Geophysical Research*, **98**(B6): 9591–9603. <https://doi.org/10.1029/93JB00716>
- Smith WHF, Sandwell DT. 1995. Marine gravity field from declassified Geosat and ERS–1 altimetry. *Eos Transactions of the American Geophysical Union*, **76**, Fall Mtng Suppl: F156.
- Smyth H, Hall R, Hamilton J, Kinny P. 2005. East Java: Cenozoic Basins, Volcanoes and Ancient Basement. *Proceedings, Indonesian Petroleum Association, Thirtieth Annual Convention & Exhibition*, August 2005.
- Socquet A, Hollingsworth J, Pathier E, Bouchon M. 2019. Evidence of supershear during the 2018 magnitude 7.5 Palu earthquake from space geodesy. *Nature Geoscience*, **12**(3): 192–199. <https://doi.org/10.1038/s41561-018-0296-0>
- Song R, Hattori K, Zhang X, Sanaka S. 2020. Seismic–ionospheric effects prior to four earthquakes in Indonesia detected by the China seismo–electromagnetic satellite. *Journal of Atmospheric and Solar–Terrestrial Physics*, **205**: 105291. <https://doi.org/10.1016/j.jastp.2020.105291>
- Sørensen M, Atakan K. 2008. Continued earthquake hazard in Northern Sumatra: Potential effects of a future earthquake. *Eos Transactions of the American Geophysical Union*, **89**(14): 133–134. <https://doi.org/10.1029/2008EO140001>
- Stagpoole V, Schenke HW, Ohara Y. 2016. A name directory for the ocean floor. *Eos*, **97**. <https://doi.org/10.1029/2016EO063177>
- Straume EO, Gaina C, Medvedev S, Hochmuth K, Gohl K, Whittaker JM, Abdul Fattah R, Doornenbal JC, Hopper JR. 2019. GlobSed: Updated total sediment thickness in the world's oceans. *Geochemistry, Geophysics, Geosystems*, **20**(4): 1756–1772. <https://doi.org/10.1029/2018GC008115>
- Suetova IA, Ushakova LA, Lemenkova P. 2005. Geoinformation mapping of the Barents and Pechora Seas. *Geography and Natural Resources*, **4**: 138–142. <https://doi.org/10.6084/m9.figshare.7435535>
- Tozer B, Sandwell DT, Smith WHF, Olson C, Beale JR, Wessel P. 2019. Global bathymetry and topography at 15 arc sec: SRTM15+. *Earth and Space Science*, **6**(10): 1847–1864. <https://doi.org/10.1029/2019EA000658>
- Tregoning P, Brunner FK, Bock Y, Puntodewo SSO, McCaffrey R, Genrich JF, Calais E, Rais J, Subarya C. 1994. First geodetic measurement of convergence

- across the Sunda Trench. *Geophysical Research Letters*, **21**(19): 2135–2138.  
<https://doi.org/10.1029/94GL01856>
- Van Zuidam RA. 1985. *Aerial Photo-Interpretation Terrain Analysis and Geomorphology Mapping*. Smits Publishers, The Hague, 442 p.
- Verstappen HT. 2000. *Outline of the Geomorphology of Indonesia: A Case Study on Tropical Geomorphology of a Tectogene Region*. ITC Publication 79, Enschede.
- Yulianto E, Utari P, Satyawati IA. 2020. Communication technology support in disaster-prone areas: Case study of earthquake, tsunami and liquefaction in Palu, Indonesia. *International Journal of Disaster Risk Reduction*, **45**: 101457.  
<https://doi.org/10.1016/j.ijdr.2019.101457>
- Weatherall P, Marks KM, Jakobsson M, Schmitt T, Tani S, Arndt JE, Rovere M, Chayes D, Ferrini V, Wigley R. 2015. A new digital bathymetric model of the world's oceans. *Earth and Space Science*, **2**(8): 331–345  
<https://doi.org/10.1002/2015EA000107>
- Wessel P, Watts AB. 1988. On the accuracy of marine gravity measurements. *Journal of Geophysical Research*, **93**: 393–413.  
<https://doi.org/10.1029/JB093iB01p00393>
- Wessel P, Smith WHF. 1991. Free software helps map and display data. *Eos Transactions of the American Geophysical Union*, **72**(41): 441.  
<https://doi.org/10.1029/90EO00319>
- Wessel P, Smith WHF. 1996. A Global Self-consistent, Hierarchical, High-resolution Shoreline Database. *Journal of Geophysical Research*, **101**: 8741–8743.  
<https://doi.org/10.1029/96JB00104>
- Wessel P, Smith WHF, Scharroo R, Luis JF, Wobbe F. 2013. Generic mapping tools: Improved version released. *Eos Transactions of the American Geophysical Union*, **94**(45): 409–410.  
<https://doi.org/10.1002/2013EO450001>
- Widiyantoro S, Pesicek JD, Thurber CH. 2011. Subducting slab structure below the eastern Sunda arc inferred from non-linear seismic tomographic imaging. *Geological Society, London, Special Publications*, **355**(1): 139–155. <https://doi.org/10.1144/SP355.7>



---

# Investigating land surface deformation using InSAR and GIS techniques in Cluj–Napoca city’s most affected sector by urban sprawl (Romania)

Anna–Hajnalka KERÉKES<sup>1\*</sup>, Szilárd Lehel POSZET<sup>2</sup>, Călin BACIU<sup>1</sup>

<sup>1</sup>Faculty of Environmental Science and Engineering, Babeş–Bolyai University, Cluj–Napoca, Romania

<sup>2</sup>Sapientia Hungarian University of Transylvania, Faculty of Sciences and Arts, Department of Environmental Sciences, Cluj–Napoca, Romania

Received 30 September 2020; Revised 20 November 2020; Accepted 23 November 2020

\*Correspondence to: Anna–Hajnalka KERÉKES; e-mail: [annakrks@yahoo.com](mailto:annakrks@yahoo.com)

---

## ABSTRACT

The last three decades have marked an unprecedented urban expansion of the city of Cluj–Napoca, leading to strong anthropogenic influences on the natural environment and important changes in the land-use. Due to the specific morphology of Cluj area, characterized by restrained plane surfaces, which are insufficient to support the urban expansion, more and more territories with slopes between 5°–26° are used for constructions. These areas are marked by high risks of mass movements due to their specific geological and geomorphological characteristics, therefore the present study proposes a more detailed and complex GIS and remote sensing analysis of Cluj–Napoca’s most affected area by urban sprawl, in order to highlight the main changes of the city and the consequences of the human actions. One of the most used radar interferometry techniques (InSAR) was applied to detect land deformations that can threaten the infrastructure and the population. Sentinel–1B SAR imagery was processed by DInSAR methodology, resulting in a land surface deformation map, which represents an important support in generating a predisposition assessment. Based upon this evaluation, we concluded that the most predisposed neighbourhoods (from the study area) to active land deformations are the peripheral ones, as following: Dâmbul Rotund, Bună Ziua, Europa, Mănăştur, West Iris and Făget, proving that human activity and the geological setting are the main triggering factors of the discussed phenomenon.

**KEYWORDS** GIS, remote sensing, Cluj–Napoca, land surface deformation, urban planning, InSAR, DInSAR

---

## 1. Introduction

Mass movement or mass wasting represents the displacement of materials (snow, ice, soil, debris,

rock) under the force of gravity, in downhill or downward direction, inducing land surface deformations (natural or anthropic hazards), like land-

sliding, subsidence, soil creeping, rockfalls, mudflows, debris flows and avalanches (Coleman and Prior, 1988; Griffiths, 2018; Chen et al., 2019). These processes can cause severe social and economic damages, therefore monitoring, mapping land surface deformations with different up-to-date Earth Observation methods and technologies becomes essential. Producing vulnerability, susceptibility and predisposition frameworks are fundamental to reduce the impact of damage (Sun et al., 2015), thus helping the improvement of urban planning for future expansion strategies.

In the past decades, many studies and investigations used remote sensing techniques in order to examine and investigate mass movement processes, revealing spatial and temporal information and data. The InSAR satellite-based technique (Interferometric Synthetic Aperture Radar) is a widespread method to measure mass movements and displacements of the Earth's surface, with millimetric accuracy (Fiaschi et al., 2017; Wempen, 2020). Gabriel et al. (1989) were the first researchers that rendered the theoretical approach of InSAR and successfully generated a map with small elevation changes. Based upon their research, Massonnet was the first who applied this technique to detect ground motion (Massonnet et al., 1993; Massonnet et al., 1995). The technology was widely and efficiently used in the past years to quantify geological hazards, like volcanic and seismic activity (Peltier et al., 2010; Antonielli et al., 2014; Chini et al., 2015) and different types of land displacements, like subsidence, debris flow, soil-creeping and landslides (Carnec et al., 1996; Galloway et al. 1998; Cascini et al., 2010; Abidin et al., 2013; Heckmann et al., 2014; Calò et al., 2014; Solari et al., 2016). Moreover, due to constant development in precision and accuracy of satellite imagery and remote sensing techniques, the InSAR method became more efficient, budget- and time-friendly (in comparison to the classical methods) to generate multitemporal measurements of large sections of areas (Cascini et al., 2010; Ullo et al., 2019). Numerous studies successfully used SAR with GIS techniques to investigate and map vulnerable urban areas to different types of terrain displacements caused by geohazards, like landsliding, subsidence and volcanic activity, taking into account

social and economic dimensions (Béjar-Pizarro et al., 2017; Solari et al., 2018; Ezquerro et al., 2020). In these studies, infrastructure damages were effectively approximated and vulnerable buildings maps were generated. Susceptibility frameworks were also produced by using radar technology, predicting highly prone areas to different mass movement types based on multiple environmental factors (Zhao et al., 2019; Bianchini et al., 2019; Jiaxuan et al., 2020).

The interferometrical SAR techniques can be grouped into two different classes: DInSAR (Differential InSAR) and PSI (Persistent Scatterers Interferometry) method. The PSI technique was created because DInSAR is only capable of measuring the differences and shifts between two points (two images) in time, it does not possess the ability to differentiate the linear motion from the nonlinear one, and it also has its limitations caused by atmospheric or signal disturbances and changes in reflections of objects (Oštir and Komac, 2007; Fiaschi et al., 2017). Therefore, the PSI approach is able to overcome the above-mentioned limitations by using highly correlated dominating scatters (Sun et al., 2015) and can also be grouped into two main classes: PSInSAR (Permanent Scatterers InSAR) and Small Baseline Subset (SBAS). PSInSAR is capable to determine, over a longer period, coherent radar targets (Oštir and Komac, 2007), whereas SBAS measures the displacements of distributed scatterers by combining radar images with small baselines (Fiaschi et al., 2017), moreover, these methods were successfully used by Ferretti et al., 2001; Lanari et al., 2004; Lauknes, 2010; Solari et al., 2016; Ma et al., 2016; Solari et al., 2018).

The DInSAR technique can be suitably applied on large areas to measure surface displacements, offers a quick look of their apportionment, generates continuous data and detects slow deformations (Oštir and Komac, 2007; Dalla Via et al., 2012; Wempen, 2020). The major advantage of the DInSAR method compared to any other interferometrical techniques lies in the actual results, which represent a continuous surface, unlike in the other techniques, where the resulted maps contain only point clouds, making them discontinuous (Oštir and Komac, 2007). The DInSAR method utilizes fewer

computing resources and only requires two images for the processing algorithm to work. Furthermore, this technology offers qualitative information on land deformations and movements due to its reduced topographic components of the phase, which are based on an external DEM, and can be generated with even fewer SAR data acquisitions (Fiaschi et al., 2017). Temporal decorrelation becomes an important issue in DInSAR processing, as it can cause difficulties when comparing radar imagery (Oštir and Komac, 2007). This major problem can be surpassed by pairing imagery over very short periods of time and by choosing high coherence areas (urban and non-vegetated, bare areas) (Oštir and Komac, 2007; Wempen, 2020). The Sentinel–1B SAR imagery is very suitable for DInSAR processing due to its very short temporal baseline (12 days) (Pawluszek–Filipiak and Borkowski, 2020), moreover, the satellite's sensor operates on C–band, which has relatively good performance and results over built-up areas and bare/rock surfaces (snow-free) (Herrera et al., 2013).

On a national scale, in Romania, the InSAR technique is relatively new and was used to detect landslide movements in Iași (Necula et al., 2017) and also to measure the subsidence rate in Bucharest (Poncos et al., 2014; Pătrașcu et al., 2016; Armaș et al., 2016; Dănișor and Dâtcu, 2018; Gheorghe et al., 2020).

In this research, social and economic dimensions were not taken into account due to the complexity and volume of data acquisition. A particular condition (in our case, the active land deformations from the past year) was taken into consideration, which can likely describe the behaviour (predisposition) of the study area qualitatively. Therefore, we did not accentuate susceptibility frameworks in this study, because our scope is to describe and quantify further possible displacements of a geomorphological event's surrounding area by using a single pre-existing condition.

Therefore, our exploratory research aims to display a potential usage of Synthetic Aperture Radar Interferometry (InSAR) and to provide an up-to-date, fast and advanced methodology to investigate Cluj–Napoca city's most active land surface deformations that have occurred in the last year and were

most affected by urban sprawl. Ground displacement in built-up areas can have a great impact on the development and land-planning strategies of cities, especially if they are human-induced (Armaș, 2017), therefore, our main objective is to gain a more profound knowledge of the geographical positions and velocities of the landmass displacements analysed between 2019 and 2020, using radar data. Also, by generating a predisposition map based on the modified methodologies of Gligor et al. (2012) and Ciampalini et al. (2016), we can describe how reliable an area can be for construction works. Moreover, by combining GIS with remote sensing techniques, we can delineate the neighbourhoods, within the study area, which are predisposed to land surface displacements.

## **2. Study area**

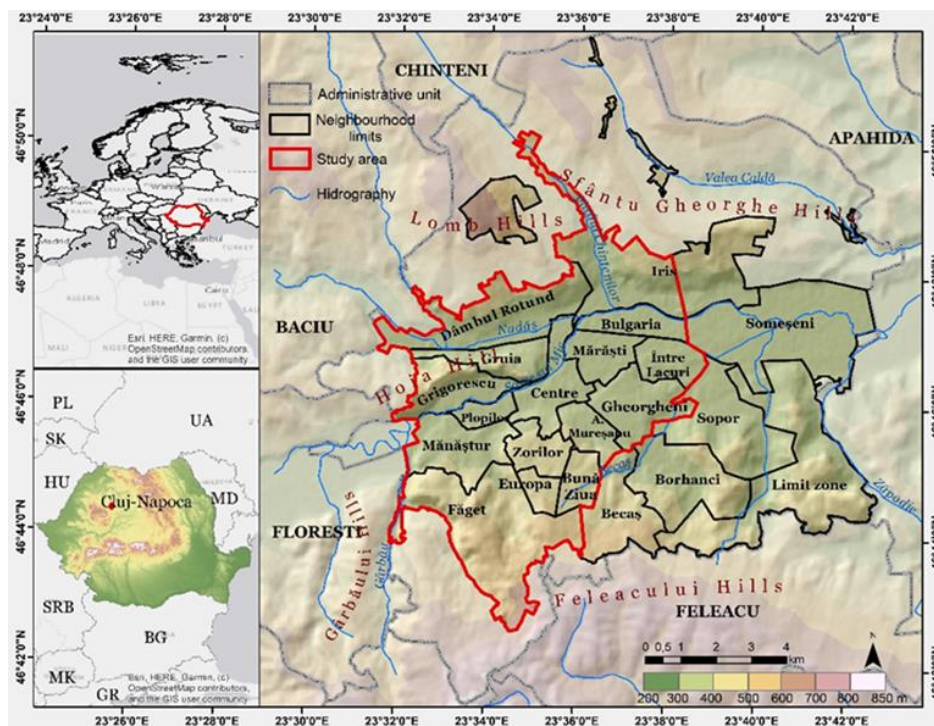
Cluj–Napoca is placed in a hilly environment in the central-western part of Transylvania, over the Someșul Mic and Nadăș River valleys, at the intersection of the 46°46'N parallel with the 23°36'E meridian. In the northern part of the municipality, the Sfântu Gheorghe and Lomb Hills are situated, the western part occupies the Hoia Hill, and the southern side includes a part of the Feleacului and Gârâului Hills (Fig. 1). In the past 30 years, the city limit has extended by 158%, therefore many constructions were built on the steeper slopes of the surrounding hills with 5°–26° inclination (Fig. 2), leading to a rapid trigger of land surface deformations. As an area of interest, we have selected a part of the city, which is the most affected by urban sprawl and built-up area extension. The geographical location of the study area is shown in Fig. 1. Moreover, in the future, we intend to conduct further DInSAR studies for the remaining part of the municipality, so we can compare the displacement values of the two distinct parts of the city described above, in order to determine the behaviour of deformations on two different geological structures. This split is also a very resource-oriented method, allowing us to use less computing power to achieve the desired accuracy of the displacement values.

The study area represents the most populated and dynamic part of the city, it incorporates the most important points of interest and has a higher

built-up rate compared to the remaining part. The territory is heavily influenced by anthropogenic and peri-urban activities, it is very dynamic in terms of buildings and road construction, land-use and land-cover changes (deforested areas, pastures and orchards turned into built-up areas), determining an increase in geological and geomorphological hazards.

The description of the geological setting of the study area was compiled after Baciú and Filipescu (2002) and Poszet (2017). The geological structure of the study area includes Paleogene (Late Eocene

and Early Oligocene) and Neogene (Early- and Middle-Miocene) sedimentary formations, partially covered by Quaternary (Pleistocene and Holocene) detrital deposits. Analysing the lithostratigraphic units (detailed in Fig. 3) and the geological map (Fig. 4) of the study area, it can be observed that the most predominant formations (Brebi, Moigrad and Dâncu, Coruş and Chechiş, Feleac, and Iris Formations), aside from the Dej Tuff complex consist of marls, clays, sands and sandstone, making the study area very susceptible to different mass movement/ land deformation processes.

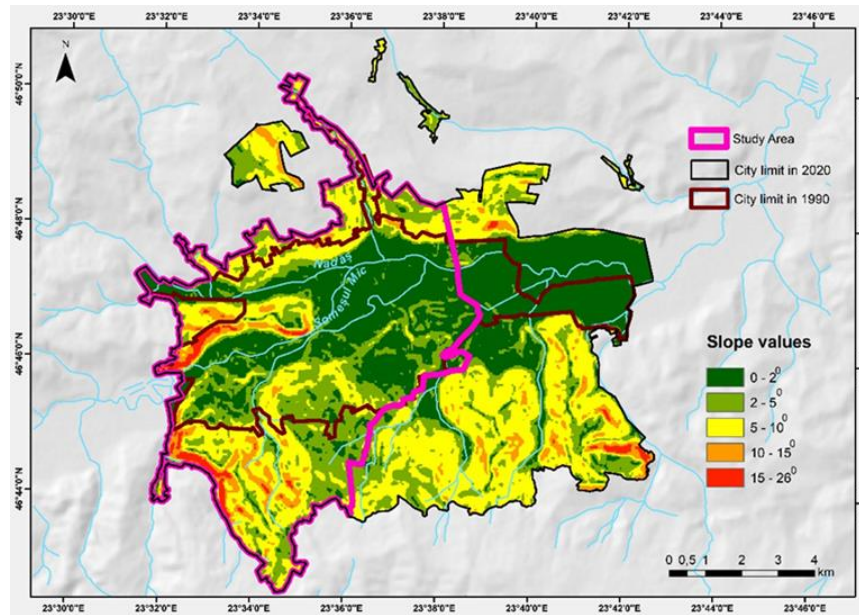


**Figure 1** The geographical position of Cluj-Napoca city (in Europe and Romania) and the extension of the study area

The architecture of the strata is generally monoclinical in the study area, and features large folds to the east. Some large faults affect the general structure, the most conspicuous of them cutting the Paleogene deposits on the eastern side of Cetățuia Hill (Fig. 4).

The geomorphological structure is highly influenced by the above-described geological formations, by the periglacial conditions of the Quaternary, river erosion and anthropogenic effects (Poszet, 2017). Fluvial and superficial deposits were formed during the Quaternary, with the seven terraces of the Someşul Mic River as the most no-

ticeable elements. The Pleistocene displacements, the periodical thawing and the recent (Holocene) and Pleistocene landslides have relocated the entire mass of sediments on lower positions of the slope. In the Holocene, the intense anthropogenic activities, like deforestation, overgrazing and construction works have triggered slope failures and thus many land mass movements have been reactivated (Kerekes et al., 2018). The majority of these mass movements are landslides (both rotational and translational), soil-creeping, land subsidence (especially near the floodplain / on calcareous formations) and earth slumps (on very steep slopes).



**Figure 2** The city limit expansion over the past 30 years and the slope values of the city (derived from 5 m resolution DEM; source: digitized elevation curves from 1:5000 topographical maps)

| Mya         | Epoch/age       | Lithostratigraphic units           | Thickness | Characteristics                                                                                                                                       |
|-------------|-----------------|------------------------------------|-----------|-------------------------------------------------------------------------------------------------------------------------------------------------------|
|             | Holocene        | Alluvium and Antropogenic Landform |           |                                                                                                                                                       |
|             | Pleistocene     | Terraces and Periglacial Formation |           |                                                                                                                                                       |
| 2.5<br>11.6 |                 |                                    |           |                                                                                                                                                       |
|             | Middle Miocene  | Feleac Sandstone Formation         | –         | Consists of alternating layers of sand and clays. Susceptible to mass movements.                                                                      |
|             |                 | Iris Marlstone Formation           | –         | Contains intercalation of marls, sands and volcanic tuff.                                                                                             |
|             | Badenian        | Dej Formation                      | 20-60 m   | Has conglomerates in the lower parts, and continues with marls, and tuff. Shows increased stability.                                                  |
| 16<br>18.4  |                 | Dej Tuff Formation                 |           |                                                                                                                                                       |
|             | Early Miocene   | Coruş Formation                    | 20-30 m   | The Chechiş Formation includes clays susceptible to produce landslides. The Coruş Formation consists of sand and sandstone with fossiliferous levels. |
| 20.5<br>~30 |                 | Chechiş Clay Formation             | 1 m       |                                                                                                                                                       |
|             | Early Oligocene | Gruia Sandstone Formation          | 20-25 m   | Characterized by the alternance of sandstone and sand.                                                                                                |
|             |                 | Dâncu Formation                    | 3 m       | Includes marls and clays, freshwater fauna.                                                                                                           |
|             |                 | Moigrad Formation                  | 60-150 m  | Consists of a pile of red clays. Very susceptible to landslide.                                                                                       |
|             |                 | Mera Formation                     | 30-35 m   | Limestone formations, commonly mapped together.                                                                                                       |
|             |                 | Hoia Limestone Formation           | ~2 m      |                                                                                                                                                       |
| 33.9        |                 | Brebi Marlstone Formation          | 50-60 m   | Contains pelite-carbonate sediments. Very rich in fossils ( <i>Bryozoans</i> ).                                                                       |
|             | Late Eocene     | Cluj Limestone Formation           | 20-50 m   | Consists of fossiliferous limestone.                                                                                                                  |
|             |                 | Jebuc Formation                    | 20-50 m   | Consists of gypsum, limestone.                                                                                                                        |
| ~37.2       |                 | Nadăş Valley Formation             | 60-80 m   | Consists of red clays, siltites and sands.                                                                                                            |

**Figure 3** The lithostratigraphic units of the study area (after Poszet, 2017)

### 3. Data and methods

The approach applied in this study consists of two main procedures: the generation of the land deformation map from the interferometric products, and the delineation of the predisposed areas. In

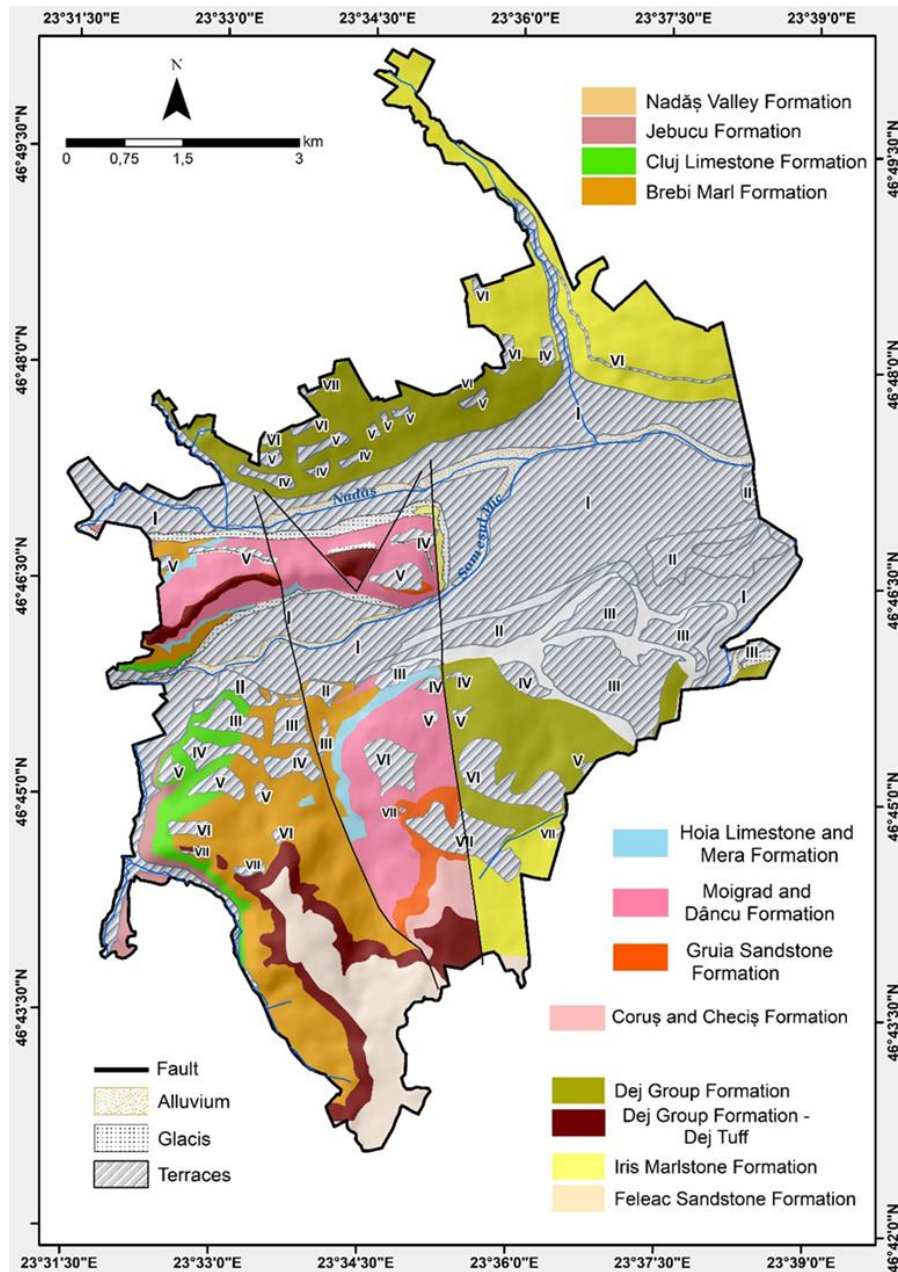
order to achieve the final maps, two specific GIS and remote sensing software products were used: SNAP 7.0 (developed by the European Spatial Agency) and ArcGIS 10.5.

### 3.1 Datasets

The deformation map was derived from Sentinel-1B SAR imagery, which characteristics can be seen in Table 1. The SAR imagery was acquired from the European Space Agency (ESA) and has a ground resolution of 5 m in the range direction and 20 m in the azimuth direction. In this study, we used 11 pairs of radar images, acquired between September 2019 and August 2020, with a descending orbit and vertical polarization. To reduce the spatial decorrelation,

we used data with a small temporal baseline (approximately 1-month interval between two SAR images).

In order to emphasize the extension of the urban sprawl of the area under investigation, we also used Landsat 5 TM from 1991.06.19 and Landsat 8 OLI from 2020.06.27, imagery acquired from the U.S. Geological Survey.



**Figure 4** The geological map of the study area (after Poszet, 2017, scale 1:5000)

**Table 1** The acquired Sentinel–1B SAR imagery characteristics

| Pairs | Date                    | Elapsed time<br>(days) | Baseline (m) | Orbits        | Average incidence<br>angle |
|-------|-------------------------|------------------------|--------------|---------------|----------------------------|
| 1     | 2019.09.04 – 2019.10.10 | 36                     | –15.00       | 17883 – 18408 | 47                         |
| 2     | 2019.10.10 – 2019.11.03 | 24                     | –19.40       | 18408 – 18758 | 47                         |
| 3     | 2019.11.03 – 2019.12.09 | 36                     | –68.90       | 18758 – 19283 | 47                         |
| 4     | 2019.12.09 – 2020.01.14 | 36                     | –36.00       | 19283 – 19808 | 47                         |
| 5     | 2020.01.14 – 2020.02.07 | 24                     | –72.28       | 19808 – 20158 | 47                         |
| 6     | 2020.02.07 – 2020.03.02 | 24                     | –187.21      | 20158 – 20508 | 47                         |
| 7     | 2020.03.02 – 2020.04.07 | 36                     | –160.12      | 20508 – 21033 | 47                         |
| 8     | 2020.04.07 – 2020.05.01 | 24                     | –130.54      | 21033 – 21383 | 47                         |
| 9     | 2020.05.01 – 2020.06.18 | 48                     | 54.06        | 21383 – 22083 | 47                         |
| 10    | 2020.06.18 – 2020.07.24 | 36                     | –3.67        | 22083 – 22608 | 47                         |
| 11    | 2020.07.24 – 2020.08.29 | 36                     | 21.18        | 22608 – 23133 | 47                         |

### 3.2. Methods

In this study, we chose to render the displacement map using DInSAR methodology owing the fact that the PSI technique utilizes advanced computer resources, thus becoming time–inefficient. Moreover, using the Differential Interferometry SAR technique, the resulted velocity map represents a continuous, grid–based surface, whereas the output of the PSI model is vector–based and shows discontinuous data. Therefore, by using the DInSAR method, data interpretation and spatial analysis becomes easier. The DInSAR methodology is suitable for dynamic process analysis (Ullo et al., 2019). It is based on measuring the phase shift between two successive SAR images which must be acquired (in different time periods) over the same location and their orbital parameters are well known (Oštir and Komac, 2007; Fiaschi et al., 2017). By processing these two SAR images, an interferometric phase (phase difference between the two images) is obtained, which is used to calculate the relative ground deformations (Fárová et al, 2019; Fiaschi et al., 2017). The interferometric phase can be described with the following formula (Hanssen, 2001, Ullo et al., 2019):

$$\Delta\phi = \Delta\phi_{\text{flat}} + \Delta\phi_{\text{height}} + \Delta\phi_{\text{displacement}} + \Delta\phi_{\text{atmosphere}} + \Delta\phi_{\text{noise}} \quad (1)$$

where  $\Delta\phi_{\text{flat}}$  represents the phase of the flat–Earth,  $\Delta\phi_{\text{height}}$  means the residual topography induced phase,  $\Delta\phi_{\text{displacement}}$  stands for the displacement measured along LOS (line of sight – contains both vertical and horizontal displacement) between the observations,  $\Delta\phi_{\text{atmosphere}}$  shows the delay of the

electromagnetic waves caused by the changes in the ionospheric dielectric constant and  $\Delta\phi_{\text{noise}}$  stands for the decorrelations/residual noise.

The DInSAR method aims to eliminate all the phase contributions mentioned above, aside from displacement, therefore obtaining the relative ground deformations (Hanssen, 2001).

Data processing was carried out based on the classical methodology described by Ullo et al., 2019, as follows:

- Image Coregistration: the SAR images must be aligned to be able to be compared pixel–by–pixel for difference detection.
- Interferogram Formation: the flat–Earth phase is removed by using orbital information and metadata.
- Multilooking and Phase Filtering (Goldstein filtering): these steps are used to remove speckle noise and it also enhances the accuracy in the course of the unwrapping step.
- Topographic phase removal: using an a priori available DEM, the topographic and geometric phase contribution can be eliminated.
- Phase Unwrapping: this step is carried out using the SNAPHU plugin; the algorithm is developed by Stanford University. Phase unwrapping resolves the ambiguity of interferometric phase (corresponding to multiples of  $|2\pi|$ ), therefore it is replacing the rectified multiples of  $|2\pi|$  of each pixel. After these steps, the LOS displacements are obtained.

This methodology was applied on the above mentioned 11 pairs of SAR images (consecutive

DInSAR) and they were all summed up in order to generate the final displacement map.

After obtaining the deformation map, we applied the approaches used by Gligor et al. (2012) and Ciampalini et al. (2016) in order to estimate, for the past year, the areas with great predisposition rates to land surface deformations. In addition to these methodologies, we introduced a change in the approximation of the predisposition rates by using weighted area values, therefore we associated the 5 LOS land displacement classes (obtained by natural break distribution) with the following weights: 0 – (0 m ÷ 0.01 m); 1 – (–0.03 m ÷ 0 m), 2 – (–0.05 m ÷ –0.03 m) and (0.01 m ÷ 0.05 m), 3 – (0.05 m ÷ 0.10 m) and (–0.07 m ÷ –0.05 m), where 0 means relatively stable surface, 1 stands for slow deformations, 2 represents medium speed mass movements and 3 stands for relatively high velocity land deformations. In order to estimate the predisposition and the sensitivity to further displacements of a geomorphological event's surrounding area, we divided the deformation map into square cells, each containing areas with different displacements and weight values. The average area weighted values formula was used in order to estimate the predisposition for each cell:

$$P = \sum_{i=1}^n (A_{di} / A_c) * W_i \quad (2)$$

where, P represents the predisposition estimation value of a cell,  $A_{di}$  represents an area from a land displacement class,  $A_c$  is equal to the cell area,  $W_i$  means the weight of the respective land displacement class and n is the number of displacement areas within the cell. The results are between 0 and 3, representing the same deformation scales presented above.

To obtain the best fit, we tested the above-mentioned approach on different cell sizes, on 100 × 100 m and 200 × 200 m, and we were able to establish different technical purposes for each scale.

The urban sprawl is an important factor which influences the land deformation; therefore, the expansion of the investigated area was emphasized using the BU (Built-up) Index based on Landsat im-

agery. This index can be calculated with the following formula (Chunyang et al., 2010):

$$BUI = NDBI - NDVI \quad (3)$$

where,

$$NDBI \text{ (Normalized Difference Building Index)} = \frac{\text{Near Infrared band} - \text{Shortwave Infrared band}}{\text{Near Infrared band} + \text{Shortwave Infrared band}} \quad (4)$$

$$NDVI \text{ (Normalized Difference Vegetation Index)} = \frac{\text{Near Infrared band} - \text{Red band}}{\text{Near Infrared band} + \text{Red band}} \quad (5)$$

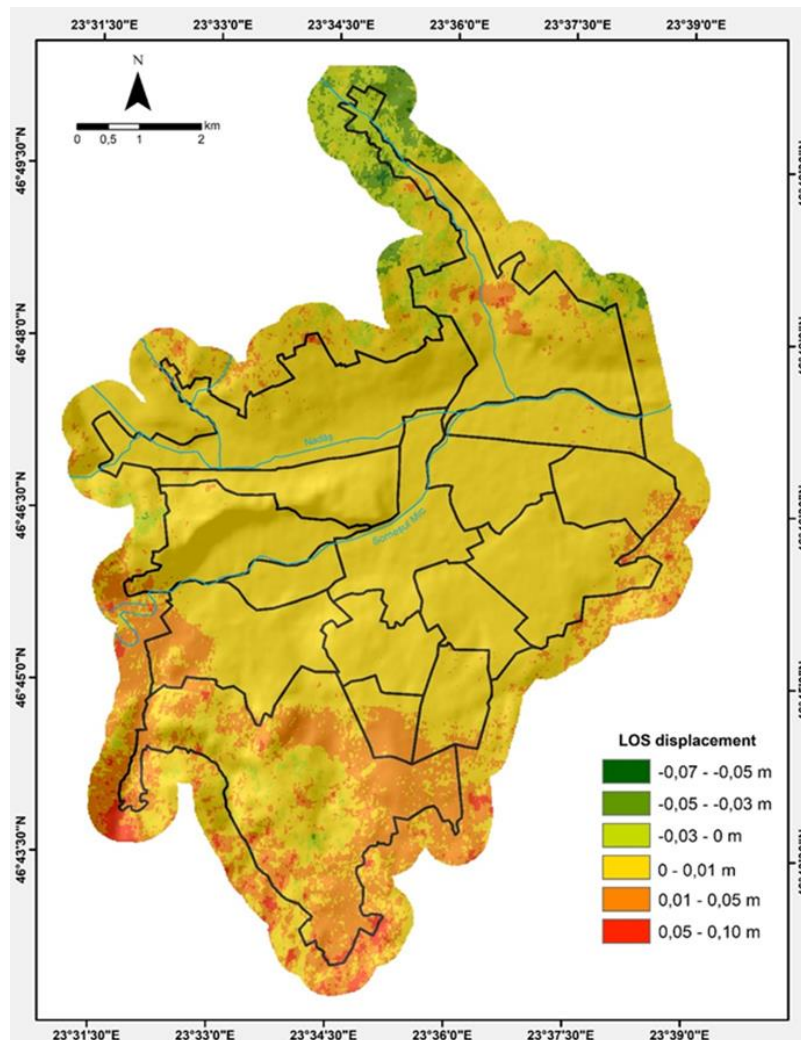
The resulted map helped us to identify different expansion tendencies and by crossing it with the predisposition estimation, we were able to calculate the percentage of the predisposed areas.

### 3.3. Validation

After generating the final predisposition map, the next step is to evaluate our estimates. Field data and measurements are not available due to the fact that many terrain deformations occurred on private properties or on inaccessible areas, therefore, we have used the field validation method.

## 4. Results and discussion

By using the consecutive DInSAR method on all the 11 pairs of SAR imagery, the perpendicular and temporal baseline values remained low, therefore avoiding temporal and spatial decorrelation. Due to this methodology, the final displacement (deformation) map (Fig. 5) is capable of showing small, millimetric deformations. The resulted map suggests that high displacement values (which highlight the magnitude of the phenomena) appear near the steeper slopes, whereas low LOS values appear on the floodplain and terraces of the Someşul Mic River, confirming that the stable part of the city is represented by the before mentioned areas. In order to understand the distribution of high LOS velocity values, the geology of the territory must be analysed together with the complex urbanisation processes.



**Figure 5** The deformation map of the study area based on DInSAR measurement

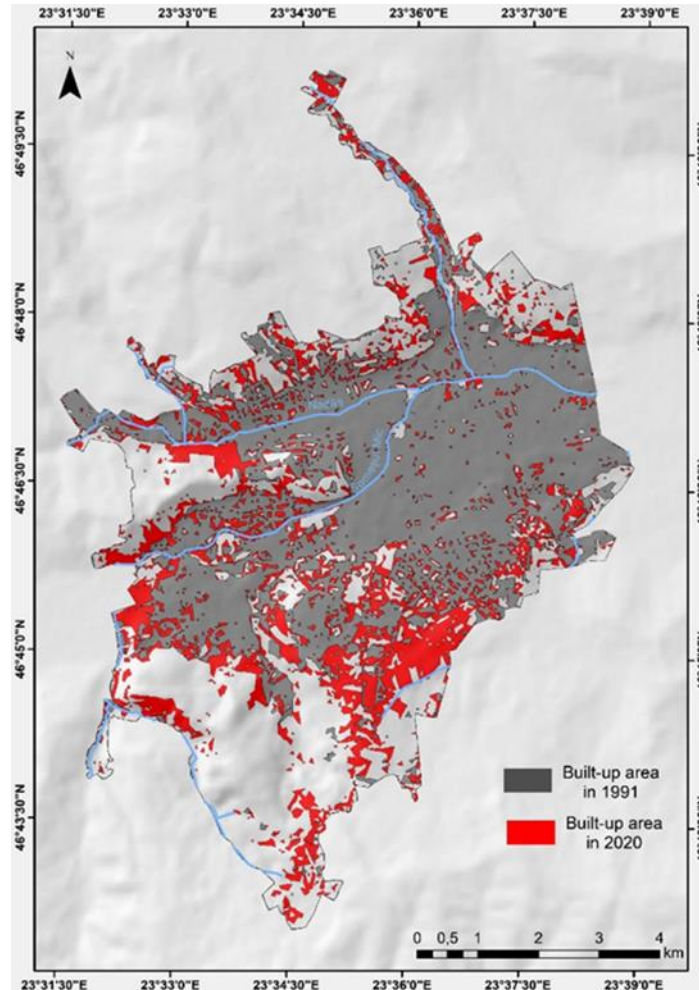
The urban sprawl phenomenon of the study area can be investigated by looking over the BU Index map (Fig. 6), therefore observing that over the last 30 years the built-up area has increased by approximately 32%. Moreover, this expansion mainly occurs in the northern (Lomb Hills), western (Hoia Hill), and southern (Feleacului Hills, Făget, Bună Ziua, Europa) parts of the city, towards areas with steeper slopes, where new constructions have been built, and the LOS values are very high (−0.07 m and 0.10 m). Moreover, these slopes are characterized by sandy–clayey formations. Thus, besides the geological and geomorphological setting, we deduced that the land deformation pattern from the last year is highly influenced by anthropogenic factors, hence

the city's built-up area tendency and expansion becomes compromised.

The aim of the predisposition map, which is generated by using the LOS velocity values, is to gain a clearer and more qualitative knowledge of the pattern and concentration of land deformations. We rendered two different predisposition maps on two different scales: 100 × 100 m and 200 × 200 m, respectively (Fig. 7). We chose these two values because they represent an average area of construction. By comparing these two maps, it can be observed that the further the scale is increased, the more the cell values are being merged to the neighbouring ones, often changing their weight towards lower values. Based on this remark, we concluded that the 200 × 200 m map offers an overall

view of the predisposition of the study area, whereas the  $100 \times 100$  m becomes more precise for technical works. It is very important to study the surroundings of a mass movement because it can be a potential disruption factor to an even bigger scale and area than itself, if the slope stability is damaged. The generated predisposition maps must become reliable, therefore field validation activity was per-

formed based on the methodology described by Righini et al. (2012). During this process, the maps described above were compared to the actual field conditions. Deteriorated buildings, damaged roads and landscapes represented hard evidence of land surface deformations in correspondence to the highest values of the predisposition map.



**Figure 6** The built-up area evolution of the study area based on the BU Index

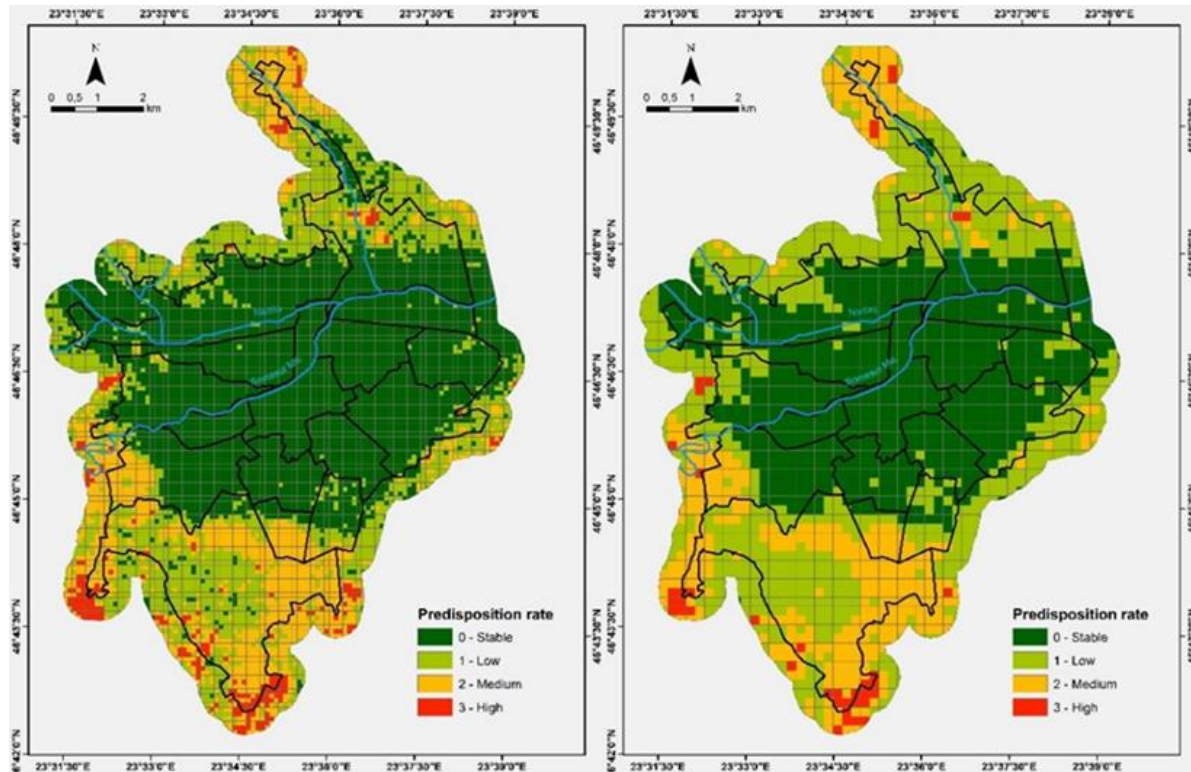
During the field survey, we observed that the southern, western and northern part of the study area is affected by landslides, soil-creeping and earth-slumps which correspond with the very high predisposition values of the map. As examples, a classical issue of mass displacement can be observed on Dragalina Str., which is located on Cetățuia–Hoia Hill (Fig. 8). A continuous 10-year movement of a landslide up to present, caused by land cover change and construction works can be

clearly observed. The mentioned location is very unsafe for construction and the reactivation of the landslide becomes a continuous and commonly known issue. A similar reactivation example can be seen on Tăietura Turcului Str. (Cetățuia–Hoia Hill) caused by the lack of terrain stability, but also influenced by excavations, vibrations and construction works (Fig. 9). Attempts to mitigate the reactivation of the mass movement have also been done, but, unfortunately, the mesh / material that was de-

signed to hold the terrain is in a constant state of degradation, presenting important risk of collapse. Severe road and construction damages appeared in multiple locations of Cluj–Napoca. An example can be seen on Fig. 10 where Uliului Str. (Grigorescu neighbourhood) was reinforced several times, but due to the local particular conditions, it is still under risk of sliding. Also, in the same area, many cracked houses were stabilized with alternative, improvised

methods, like iron fences or concrete structures (Fig. 11), but unfortunately, new fractures still appear every year.

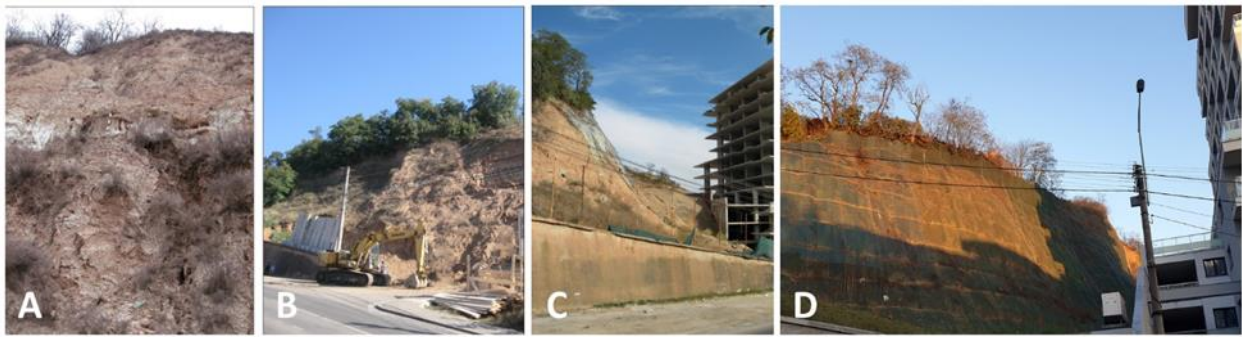
Therefore, throughout the field validation, we could observe a prominent correspondence between high terrain deformation predisposition values and the cracks on the buildings and roads, which indicates a very good reliability of the generated map.



**Figure 7** The study area's predisposition map on different scales (Left: 100 × 100 m; Right: 200 × 200 m)



**Figure 8** The reactivation of mass movements on Dragalina Str., Grigorescu neighbourhood (A: the area under dry conditions; B: mass movement reactivation due to constructions and high water content in soil; C and D: the continuous mass movement) (source of Fig. 8A and 8C: Poszet, 2017)



**Figure 9** Human induced land deformation on Tăietura Turcului (Hoia Hill), between Grigorescu and Dâmbul Rotund neighbourhoods (A: slope before excavations; B: the excavation process has disrupted the slope stability; C: concrete structures have been built in order to block the mass movements; D: the material that was designed to hold the terrain is degrading)



**Figure 10** The restoration of Uliului Str., Grigorescu neighbourhood (Left and middle: large fractures on the street; Right: the consolidated road which is still under risk of sliding)



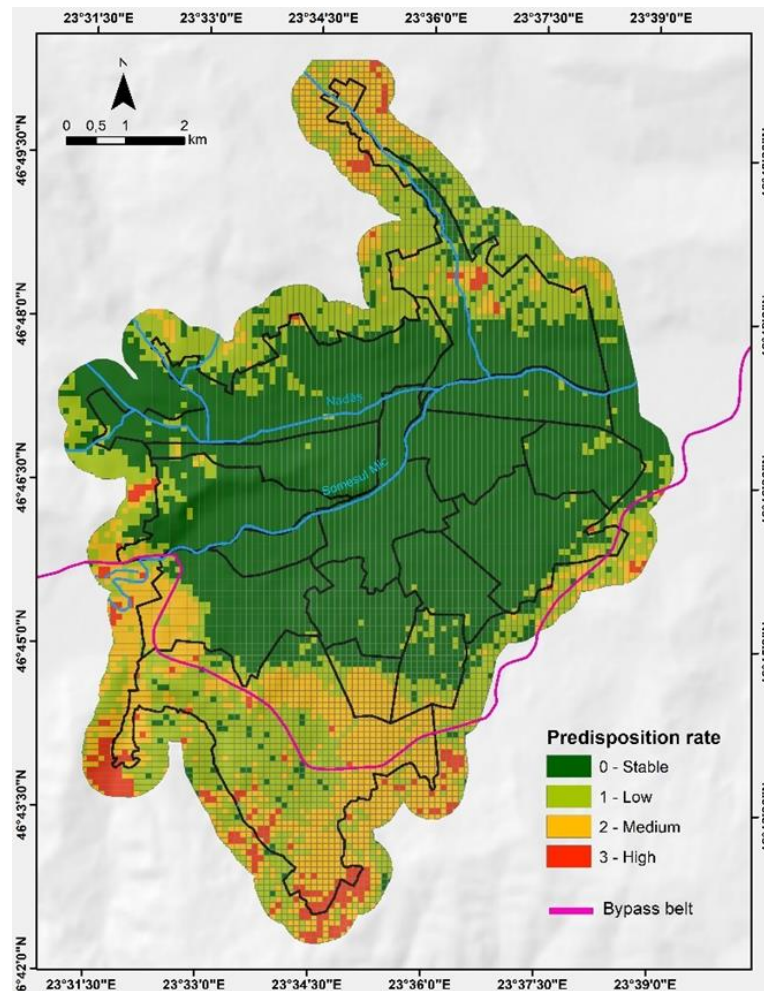
**Figure 11** Attempts to repair cracked buildings (Left: alternative concrete structures used to stabilize the building; Right: improvised methods used to hold a house together)

After the validation process, some calculations based upon the  $100 \times 100$  m map were rendered. By crossing the predisposition map with the total built-up area, we concluded that 82.42% of the buildings are constructed on safe areas, but 2.15%

of the remaining buildings are characterized by high, 7.95% by moderate and 7.48% by low predisposition rates. With the same methodology, we could calculate the percentual contribution of each predisposition class for each neighbourhood, as shown in Table 2.

**Table 2** Percentual contribution of each displacement class

| Neighbourhood   | No predisposition(0)(kmp) | Low (1) (kmp) | Moderate (2) (kmp) | High (3) (kmp) | Total area (kmp) |
|-----------------|---------------------------|---------------|--------------------|----------------|------------------|
| Andrei Mureşanu | 1.91 (97.8%)              | 0.04(2.2%)    | 0                  | 0              | 1.95             |
| Bulgaria        | 2.05 (98.08%)             | 0.04(1.92%)   | 0                  | 0              | 2.09             |
| Bună Ziua       | 0.98 (59.03%)             | 0.48 (28.92%) | 0.20 (12.05%)      | 0              | 1.66             |
| Centru          | 0                         | 0             | 0                  | 0              | 3.42             |
| Dâmbul Rotund   | 6.11 (87.80%)             | 0.82 (11.78%) | 0.02 (0.28%)       | 0.01 (0.14%)   | 6.96             |
| Europa          | 0.33 (30.84%)             | 0.15 (14.02%) | 0.59 (55.14%)      | 0              | 1.07             |
| Făget           | 0.75 (6.72%)              | 4.73 (42.43%) | 5.12(45.92%)       | 0.55 (4.93%)   | 11.15            |
| Gheorgheni      | 3.51(89.20%)              | 0.38 (9.66%)  | 0.04(1.02%)        | 0.005 (0.12%)  | 3.935            |
| Grigorescu      | 2.95 (84.94%)             | 0.43 (12.38%) | 0.09 (2.60%)       | 0.003(0.08%)   | 3.473            |
| Gruia           | 2.46 (97.23%)             | 0.05 (1.98%)  | 0.02 (0.79%)       | 0              | 2.53             |
| Iris V          | 3.42 (50.60%)             | 2.33 (34.47%) | 0.88(13.01%)       | 0.13 (1.92%)   | 6.76             |
| Între Lacuri    | 1.97 (93.80%)             | 0.13(6.20%)   | 0                  | 0              | 2.10             |
| Mănăştur        | 3.008 (70.48%)            | 0.44 (10.31%) | 0.82 (19.21%)      | 0              | 4.268            |
| Mărăşti         | 2.38 (99.96%)             | 0.0008(0.04%) | 0                  | 0              | 2.3808           |
| Plopilor        | 0                         | 0             | 0                  | 0              | 1.10             |
| Zorilor         | 2.07 (98.57%)             | 0.03(1.43%)   | 0                  | 0              | 2.10             |



**Figure 12** The path of the city's future bypass belt

In order to determine a predisposition ranking between the neighbourhoods, we calculated a weighted average from each neighbourhood's percentage contribution (the weights and the percentage contribution can be seen in Table 2). The results indicate that the most stable parts of the city are the Centre and Plopilor neighbourhood. This can be explained by the fact that these two areas are almost entirely built on the lower terraces and the floodplain of Someșul Mic River. The moderately predisposed areas are the following: Mărăști, Zorilor, Gruia, Andrei Mureșanu, Bulgaria, Gheorgheni, Grigorescu, Între Lacuri. This sequence is relatively correct, a significant part of these territories was built on terraces, but the remaining parts of them are characterized by steeper slopes, as for example Zorilor, Gruia and Grigorescu neighbourhoods. Între Lacuri and Gheorgheni neighbourhoods show land deformations near some lakes. Also, the Eastern part of the Gheorgheni neighbourhood is affected by new construction works, which explains the presence of the sensitive areas. The most predisposed neighbourhoods, Dâmbul Rotund, Bună Ziua, Europa, Mănăstur, West Iris and Făget are showing very steep slopes, intensive anthropogenic influences and land-use change. The geological constitution of these terrains, mainly rep-

resented by an alternation of sands, marls, and clays, also increases the predisposition to mass movements. Therefore, a detailed analysis of the geological structure, combined with the geomorphological features is important in understanding the reliability of an area for construction works in order to propose appropriate measures to avoid land surface deformations and their negative effects.

Due to the particular geologic and geomorphic features of the city of Cluj-Napoca, it is more and more difficult to find convenient locations and safe areas to build on. Very often, new major infrastructure projects need to cross zones predisposed to mass movements. One example is the new bypass belt project (Fig.12), which is meant to solve the road traffic problem within the urban area, and to ease the access to different parts of the city (PCJ, 2019). This is a technical and land planning strategy example in which our estimation map can be applied. According to our interpretation, the predisposition of this road to mass movements will be high for 0.42% of the road length, moderate for 39.89%, low on 51.17%, and 8.52% will be constructed on stable areas. Therefore, special technical measures should be taken in difficult sections of the route, to avoid failure.

## 5. Conclusions

The InSAR technique represents a useful method to measure terrain deformations, with millimetric accuracy and also presents a wide range of applicability in different domains, like engineering and urban planning.

In this study, the DInSAR model was applied on 11 pairs of SAR images to identify and map the land surface displacements which occurred in the last year in the urban and peri-urban area of Cluj-Napoca. The unorganized expansion of a city caused important changes in the landscape, leading to numerous mass movements, mainly caused by construction works, excavations, vibrations and land-use and cover change. By using this consecutive Differential Interferometry SAR method, we avoided decorrelation and became capable to reveal small movements that cannot be observed by simple observation. The DInSAR is a cost-effective method

which can yield important spatial and temporal information about the pattern of land deformations. Moreover, the resulted displacement map showed a continuous surface, therefore data interpretation and spatial analysis became easier.

Based upon an a priori knowledge of the study area, our generated predisposition map presents a good approximation and shows that due to the anthropogenic activities and geological characteristics, steeper slopes become more predisposed and have higher LOS rates ( $-0.07$  m and  $0.10$  m) than the plane surfaces. The approach also determines that the most predisposed neighbourhoods in Cluj-Napoca are: Dâmbul Rotund, Bună Ziua, Europa, Mănăstur, West Iris and Făget. Therefore, this model shows promising results in infrastructure monitoring. The methodology also becomes suitable for a constant surveillance of areas conditioned by human induced activities in order to prevent calamities.

ties. The predisposition map highlighted the importance of analysing the surrounding area of a mass movement and it can stand as an operational tool for construction and geotechnical plans.

Future monitoring of the study area becomes an important issue. Examining and detecting land-use changes and deforestation with remote sensing techniques can also reduce potential mass movements. Also, professional consolidation of the affected constructions is needed to prevent the pre-existing situation from getting worse and also, overloading must be avoided not to damage the slope's stability. Local authorities should have a close collaboration with mass movement experts, geotechnical engineers, and remote sensing specialists in order to render complex and reliable feasibility assessments.

Future works will be carried out on the whole municipality of Cluj–Napoca, using a greater amount of radar satellite imagery to generate further vulnerability, risk and susceptibility maps, which can stand as a support for future disaster mitigation works.

## References

- Abidin HZ, Andreas H, Gumilar I, Sidiq TP, Fukuda Y. 2013. Land subsidence in coastal city of Semarang (Indonesia): characteristics, impacts and causes. *Geomatics, Natural Hazards and Risk*, **4**(3): 226–240
- Antonielli B, Monserrat O, Bonini M, Righini G, Sani F, Luzi G, Feyzullayev AA, Aliyev CS. 2014. Pre-eruptive ground deformation of Azerbaijan mud volcanoes detected through satellite radar interferometry (DInSAR). *Tectonophysics*, **637**: 163–177
- Armaş I, Gheorghe M, Lendvai AM, Dumitru PD, Bădescu O, Călin A. 2016. InSAR validation based on GNSS measurements in Bucharest. *International Journal of Remote Sensing*, **37**(23): 5565–5580
- Armaş I, Mendes D, Popa RG, Gheorghe M, Popovici D. 2017. Long-term ground deformation patterns of Bucharest using multitemporal InSAR and multivariate dynamic analyses: A possible transpressional system? *Scientific Reports*, **7**: 43762
- Baciu C, Filipescu S. 2002. Structura geologică. In: Cristea V, Baciu C, Gafta D. (eds.) *Municipiul Cluj–Napoca și zona periurbană. Studii ambientale*. Editura Accent, Cluj–Napoca, 25–36
- Béjar–Pizarro M, Notti D, Mateos RM, Ezquerro P, Centolanza G, Herrera G, Bru G, Sanabria M, Solari L, Duro J, Fernández J. 2017. Mapping Vulnerable Urban Areas Affected by Slow–Moving Landslides Using Sentinel–1 InSAR Data. *Remote Sens.*, **9**(9), 876
- Bianchini S, Solari L, Del Soldato M, Raspini F, Montalti R, Ciampalini A, Casagli N. 2019. Ground Subsidence Susceptibility (GSS) Mapping in Grosseto Plain (Tuscany, Italy) Based on Satellite InSAR Data Using Frequency Ratio and Fuzzy Logic. *Remote Sens.*, **11**, 2015
- Calò F, Ardizzone F, Castaldo R, Lollino P, Tizzani P, Guzzetti F, Lanari R, Angeli MG, Pontoni F, Manunta M. 2014. Enhanced landslide investigations through advanced DInSAR techniques: The Ivancich case study, Assisi, Italy. *Remote Sensing of Environment*, **142**: 69–82
- Carnec C, Massonnet D, King C. 1996. Two examples of the use of SAR interferometry on displacement fields of small spatial extent. *Geophys. Res. Lett.*, **23**(24): 3579–3582
- Cascini L, Fornaro G, Peduto D. 2010. Advanced low- and full-resolution DInSAR map generation for slow-moving landslide analysis at different scales. *Engineering Geology*, **112**(1–4): 29–42
- Chen Y, Tan K, Yan S, Zhang K, Zhang H, Liu X, Li H, Sun Y. 2019. Monitoring Land Surface Displacement over Xuzhou (China) in 2015–2018 through PCA–Based Correction Applied to SAR Interferometry. *Remote Sens.*, **11**(12), 1494
- Chini M, Albano M, Saroli M, Pulvirenti L, Moro M, Bignami C, Falcucci E, Gori S, Modoni G, Pierdicca N, Stramondo S. 2015. Coseismic Liquefaction Phenomenon Analysis by Cosmo–Skymed: 2012 Emilia (Italy) Earthquake. *International Journal of Applied Earth Observation and Geoinformation*, **39**: 65–78
- Chunyang H, Peijun S, Dingyong X, Yuanyuan Z. 2010. Improving the normalized difference built-up index to map urban built-up areas using a semiautomatic segmentation approach. *Remote Sensing Letters*, **1**(4): 213–221
- Ciampalini A, Raspini F, Lagomarsino D, Catani F, Casagli N. 2016. Landslide susceptibility map refinement using PSInSAR data. *Remote Sensing of Environment*, **184**: 302–315
- Coleman JM, Prior DB. 1988. Mass wasting on continental margins. *Ann. Rev. Earth Planet. Sci.*, **16**: 101–19
- Dalla Via G, Crosetto M, Crippa B. 2012. Resolving vertical and east–west horizontal motion from differential interferometric synthetic aperture radar: the L'Aquila earthquake. *J. Geophys. Res.: Solid Earth*, **117**(B2)
- Dănișor C, Dâtcu M. 2018. Extension of PS–InSAR Approach for DEM and Linear Deformation Rates

- Estimation. Case Study of Bucharest Area. *EUSAR 2018; 12th European Conference on Synthetic Aperture Radar*, Aachen, Germany, 1–6
- Ezquerro P, Del Soldato M, Solari L, Tomás R, Raspini F, Ceccatelli M, Fernández–Merodo JA, Casagli N, Herrera G. 2020. Vulnerability Assessment of Buildings due to Land Subsidence Using InSAR Data in the Ancient Historical City of Pistoia (Italy). *Sensors*, **20**, 2749
- Fárová K, Jelének J, Kopačková–Strnadová V, Kycl P. 2019. Comparing DInSAR and PSI Techniques Employed to Sentinel–1 Data to Monitor Highway Stability: A Case Study of a Massive Dobkovický Landslide, Czech Republic. *Remote Sensing*, **11**(22), 2670
- Ferretti A, Prati C, Rocca F. 2001. Permanent scatterers in SAR interferometry. *IEEE Trans Geosci Remot Sens*, DOI: 10.1109/36.898661
- Fiaschi S, Tessitore S, Boni R, Martire D, Achilli V, Bogstrom S, Ibrahim A, Floris M, Meisina C, Ramondini M, Calcaterra D. 2017. From ERS–1/2 to Sentinel–1: two decades of subsidence monitored through A–DInSAR techniques in the Ravenna area (Italy). *GIScience & Remote Sensing*, **54** (3)
- Gabriel K, Goldstein RM, Zebker HA. 1989. Mapping Small Elevation Changes over Large Areas: Differential Radar Interferometry. *Journal of Geophysical Research*, **94**(B7): 9183–9191
- Galloway DL, Hudnut KW, Ingebritsen S, Philips SP, Peltzer G, Rogez F, Rosen PA. 1998. Detection of aquifer system compaction and land subsidence using interferometric synthetic aperture radar, Antelope Valley, Mojave Desert. *California Water Resour. Res.*, **34** (10): 2573–2585
- Gheorghe M, Armaş I, Dumitru P, Călin A, Bădescu O, Necşoiu M. 2020. Monitoring subway construction using Sentinel–1 data: a case study in Bucharest. Romania. *International Journal of Remote Sensing*, **41**(7): 2644–266
- Gligor V, Bilaşco Ş, Fonogea SF. 2012. Analiza morfo-funcţională a teritoriului comunei Floreşti (Judeţul Cluj) şi aprecierea gradului de vulnerabilitate locală la procesele geomorfologice de risc. *Geographia Napocensis*, **VI**(1): 11–20
- Griffiths JS. 2018. Mass Movement. In: Bobrowsky PT, Marker B. (eds.) *Encyclopedia of Engineering Geology*. Encyclopedia of Earth Sciences Series, Springer, Cham.
- Hanssen RF. 2001. *Radar Interferometry: Data Interpretation and Error Analysis*. Kluwer Academic; Springer: Dordrecht, The Netherlands, DOI: 10.1007/0-306-47633-9
- Heckmann T, Gegg K, Becht M. 2014. Sample size matters: investigating the effect of sample size on a logistic regression susceptibility model for debris flow. *Nat. Hazards Earth Syst. Sci.*, **14**: 259–278
- Herrera G, Guitierrez F, García–Davalillo JC, Guerrero J, Notti D, Galve JP, Fernández–Merodo, JA, Cooksley G. 2013. Multi–sensor advanced DInSAR monitoring of very slow landslides: The Tena Valley case study (Central Spanish Pyrenees). *Remote Sensing of Environment*, **128**: 31–43
- Jiaxuan H, Mowen X, Atkinson PM. 2020. Dynamic susceptibility mapping of slow–moving landslides using PSInSAR. *International Journal of Remote Sensing*, **41**(19): 7509–7529
- Kerekes AH, Poszet SzL, Gál A. 2018. Landslide susceptibility assessment using the maximum entropy model in a sector of the Cluj–Napoca Municipality, Romania. *Revista de Geomorfologie*, **20**: 130–146
- Lanari R, Mora O, Manunta M, Mallorqui JJ, Berardino P, Sansosti E. 2004. A small–baseline approach for investigating deformations on full–resolution differential SAR interferograms. *IEEE Transactions on Geoscience and Remote Sensing*, **42**(7): 1377–1386
- Lauknes TR, Piyush Shanker A, Dehls JF, Henderson IHC, Larsen Y. 2010. Detailed rockslide mapping in northern Norway with small baseline and persistent scatterer interferometric SAR time series methods. *Remote Sensing of Environment*, **114**: 2097–2109
- Ma C, Cheng X, Yang Y, Zhang X, Guo Z, Zou Y. 2016. Investigation on Mining Subsidence Based on Multi–Temporal InSAR and Time–Series Analysis of the Small Baseline Subset– Case Study of Working Faces 22201–1/2 in Bu’ertai Mine, Shendong Coalfield, China. *Remote Sens.*, **8**(11), 951
- Massonnet D, Rossi M, Carmona C, Adragna F, Peltzer G, Feigl K, Rabaute T. 1993. The displacement field of the Landers earthquake mapped by radar interferometry. *Nature*, **364**: 138–142
- Massonnet D, Briole P, Arnaud A. 1995. Deflation of Mount Etna monitored by spaceborne radar interferometry. *Nature*, **375**: 567–570
- Necula N, Niculiţă M, Tessari G, Floris M. 2017. InSAR analysis of Sentinel–1 data for monitoring landslide displacement of the north–eastern Copou hillslope, Iaşi city, Romania. *Proceedings SNG*, DOI 10.15551/prgs.2017.85
- Oštir K, Komac M. 2007. PSInSAR and DInSAR methodology comparison and their applicability in the field of surface deformations – A case of NW Slovenia. *Geologija*, **50**(1): 77–96
- Pawluszek–Filipiak K, Borkowski A. 2020. Comparison of PSI and DInSAR approach for the subsidence monitoring caused by coal mining exploitation. *The International Archives of the Photogrammetry*,

- Remote Sensing and Spatial Information Sciences*, Volume XLIII-B3-2020, XXIV ISPRS Congress, 333–337
- Pătraşcu C, Popescu AA, Dâtcu M. 2016. Comparative assessment of multi-temporal InSAR techniques for generation of displacement maps: a case study for Bucharest area. *U.P.B. Sci. Bull., Series C* **78**(2):135–146
- PCJ (Primăria Cluj–Napoca). 2019. Varianta finală de traseu V8 pentru centura metropolitană. <https://primariaclužnapoca.ro/informatii-publice/comunicate/varianta-finala-de-traseu-v8-pentru-centura-metropolitana/>. Accessed 10 August 2020
- Peltier A, Bianchi M, Kaminski E, Komorowski JC, Rucci A, Staudacher T. 2010. PSInSAR as a new tool to monitor pre-eruptive volcano ground deformation: Validation using GPS measurements on Piton de la Fournaise. *Geophys Res Lett.*, 37, DOI: 10.1029/2010GL043846
- Poncos V, Teleaga D, Boukhemacha MA, Toma SA, Serban F. 2014. Study of urban instability phenomena in Bucharest city based on Ps-InSAR. *IEEE Geoscience and Remote Sensing Symposium, Quebec City, QC*, 429–432
- Poszet SzL. 2017. Studiu de geomorfologie aplicată în zona urbană Cluj–Napoca, *Editura Scientia*, Cluj–Napoca.
- Righini G, Pancioli V, Casagli N. 2012. Updating landslide inventory maps using Persistent Scatterer Interferometry (PSI). *International Journal of Remote Sensing*, **33**(7): 2068–2096
- Solari L, Ciampalini A, Raspini F, Bianchini S, Moretti S. 2016. PSInSAR analysis in the Pisa Urban Area (Italy): A case study of subsidence related to stratigraphical factors and urbanization. *Remote Sens.*, **8**(120)
- Solari L, Barra A, Herrera G, Bianchini S, Monserrat O, Pizarro–Béjar M, Crosetto M, Sarro R, Moretti S. 2018. Fast detection of ground motions on vulnerable elements using Sentinel–1 InSAR data. *Geomatics, Natural Hazards and Risk*, **9**(1): 152–174
- Sun Q, Zhang L, Ding XL, Hu J, Li ZW, Zhu JJ. 2015. Slope deformation prior to Zhouqu, China landslide from InSAR time series analysis. *Remote Sensing of Environment*, **156**: 45–57
- Ullo SL, Addabbo P, Di Martire D, Sica S, Fiscante N, Cicala L, Angelino CV. 2019. Application of DInSAR Technique to High Coherence Sentinel–1 Images for Dam Monitoring and Result Validation Through In Situ Measurements. *IEEE Journal of Selected Topics in Applied Earth Observations and Remote Sensing*, **12**(3), 1939–1404
- Wempen JM. 2020. Application of DInSAR for short period monitoring of initial subsidence due to longwall mining in the mountain west United States. *International Journal of Mining Science and Technology*, **30**(1): 33–37
- Zhao F, Meng X, Zhang Y, Chen G, Su X, Yue D. 2019. Landslide Susceptibility Mapping of Karakorum Highway Combined with the Application of SBAS–InSAR Technology. *Sensors*, **19**(12): 2685



---

# Rock glaciers in mixed lithology: a case study from Northern Pirin

Emil GACHEV<sup>1,2\*</sup>

<sup>1</sup>South–West University “Neofit Rilski”, Blagoevgrad, Bulgaria

<sup>2</sup>National Institute of Geophysics, Geodesy and Geography, Bulgarian Academy of Sciences, Sofia, Bulgaria

Received 14 November 2020; Revised 16 December 2020; Accepted 16 December 2020

\*Correspondence to: Emil GACHEV; e-mail: [emil.gachev@swu.bg](mailto:emil.gachev@swu.bg)

---

## ABSTRACT

Rock glaciers are lobate or tongue-shaped assemblages of poorly sorted, angular–rock debris and ice, commonly found in high mountain environments, which move as a consequence of the deformation of internal ice (Giardino and Vitek, 1988; Barsch, 1996). However, in most research works, when discussing the formation of these features, the focus has been mainly on the past or present climatic conditions. The systematic observations of the distribution of relict rock glaciers in the mountains of the Balkan Peninsula indicate however that geological setting, represented by bedrock type and composition, and the pattern of fault lines, is not less important for the formation and evolution of these landforms.

The present article is focused on rock glaciers in the high mountain zone of the Pirin Mounains (Bulgaria), which are formed in mixed lithology, with participation of both silicate rocks (granite, gneiss) and carbonate rocks (marble). In fact, these rock glaciers are the only ones of a typical morphology that exist in the glaciokarstic marble area of Northern Pirin. What is common for all locations where such rock glaciers are found is that the marble layer, which is on top, is quite thin (few metres to few hundreds of metres). In such conditions, the rock glacier formation occurred following the mechanism typical for silicate rocks, but using marble debris and block material instead. The observed forms are characteristic only for marginal zones, along the contact line between silicate and carbonate high mountain environments.

**KEYWORDS** Pirin, rock glaciers, mixed lithology, marble, silicate rocks

---

## 1. Introduction

Rock glaciers are among the most prominent cryogenic landforms in the mountains of Southeast Europe. These are usually large deposits of rock debris of various sizes, often in form of a chaotic pattern of

ridges, furrows and lobes. In Bulgaria, rock glaciers are observed in high mountain areas, at altitudes between 2096 and 2710 m a.s.l. (Onaca et al., 2020). According to Washburn (1979) a rock glacier is “a tongue–like or lobate body usually of angular boulders that resembles a small glacier, generally occurs

in high mountainous terrain, and usually has ridges, furrows, and sometimes lobes on its surface, and has a steep front at the angle of repose". Berthling (2011) defined rock glaciers as "the visible expression of cumulative deformation by long-term creep of ice/debris mixtures under permafrost conditions". It is considered that they originate either from true glaciers (Ackert, 1998; Harrison et al., 2008), which during recession and deglaciation in warming conditions were gradually covered by debris (i. e. during warming), or from scree deposits (Haeberli, 1985), in which permafrost developed during a subsequent cooling stage.

In general, rock glaciers have been categorized as either active, inactive or relict (Wahrhaftig and Cox, 1959). Active rock glaciers have a core of buried ice inside the debris. This makes them able to move downslope, forced by gravity. The heavy weight of the debris material exerts pressure on the ice and makes it plastic, and initiates the creep process. Similar to glaciers, rock glaciers are able to move on surfaces tilted 8–10 degrees. Therefore, the debris material can travel several hundreds of metres down and away from debris' source zone. In conditions of consequent warming/decrease of solid precipitation, rock glaciers may become inactive – they still have some ice inside, but no more move. If further warming occurs, ice can melt completely, and cause some subsidence on the surface of the rock glacier, which becomes relict. Rock glaciers in the high mountains of the Carpathian–Balkan area are considered mostly relict, but for 10 of the highest rock glaciers in Rila and Pirin Mountains still the presence of sporadic patches of permafrost is suggested (Onaca et al., 2020).

Since the pioneer work of Capps (1910), numerous studies of rock glaciers have been published (Wahrhaftig and Cox, 1959; Martin and Whalley, 1987; Barsch, 1988, 1997). In most of them, rock glaciers are researched as products of climate changes in the past, and as landforms which can successfully be used as proxies for paleoclimate reconstructions (Martin and Whalley, 1987). At the same time, less attention has been paid to the importance of topography, and especially of geological setting for rock glacier formation.

Along with appropriate climatic and topographic conditions, in order to develop, rock glaciers require a source of material to supply debris with a proper size (Martin and Whalley, 1987). Prevalent rock fragments need to be large enough to provide sufficient internal space (caverns, pores) for the snow accumulation and freezing of water. The possibility to produce such fragments depends on the structure of rocks and their way of weathering. Rocks which are not highly prone to physical weathering (such as limestone and dolomite), or have thin schistosity and disintegrate into tiles (for example some schists, phyllite) do not support formation of buried ice, as the screes produced in result of their weathering are very compact and provide too little pore space. Also inappropriate are all chemically highly soluble rocks, such as the mentioned limestone and dolomite, and also marble (despite that the latter is a crystalline rock which fragments into larger blocks than limestone), because of the development of karst features. Atmospheric waters that infiltrate in the scree accumulations of such rocks drain in the karst caverns and cannot be retained in the scree long enough to form ice (even if pores are large enough).

On the contrary, among the rocks which are especially supportive for rock glacier formation, are granite, gneiss, massive conglomerate, which naturally have lots of cracks, and fragment in relatively large and angular blocks.

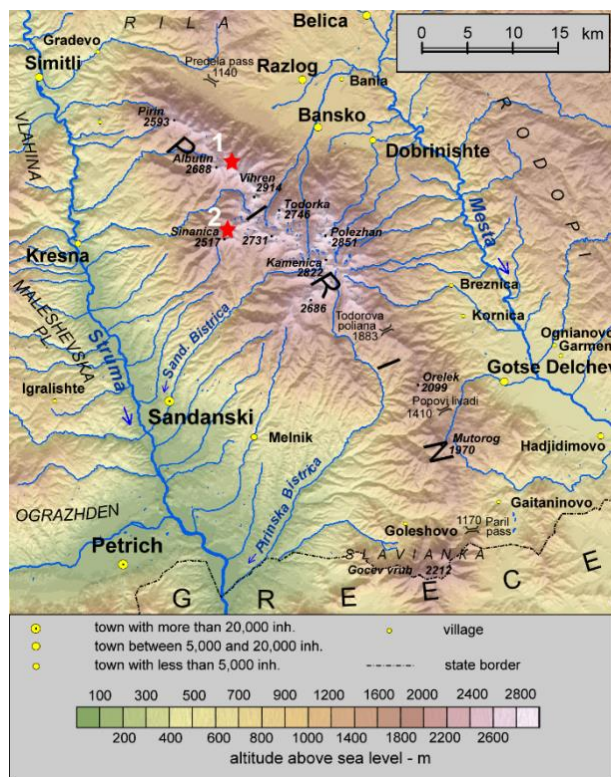
The present article focuses on some specific cases of relict rock glaciers in the Pirin Mountains of Bulgaria, which blocky substrate is composed of marble. Contrary to limestone from which it descends, marble has a tendency to develop cracks at right angles, and defragment on larger angular blocks due to physical weathering, especially frost action) and because of this it is able to produce deposits for rock glacier formation. On the other hand, marble is a carbonate rock, chemically soluble in water. Usually marble surfaces are karstified, which results in a lack of surface runoff, as atmospheric (rain and snow) waters infiltrate in karstic caverns.

The studied rock glaciers are formed in conditions of mixed lithology (marble and granite/gneiss), and have distinct morphology and rock composition. Such forms can be considered unusual, being

determined both by the diverse bedrock with contrast chemistry (carbonate vs. silicate rocks) and by the appropriate rock stratification and sequences.

## 2. Study area

Pirin is the second highest mountain in Bulgaria and the third highest on the Balkans. It is located in southwest Bulgaria, between the rivers Struma, to the west and Mesta, to the east. The mountain is generally shaped as a rhomboid (70 x 35 km size) with its longer axis running in NNW–SSE direction (Fig. 1). The highest point, Mt. Vihren (2914 m a.s.l.), is situated in the northern part of the mountain. Two more locations in the central section exceed 2800 m a.s.l., and there are numerous peaks above 2500 m a.s.l.



**Figure 1** Map of Pirin Mountains with the study objects: 1 – Razlozhi suhodol rock glacier; 2 – Sinanitsa rock glacier

Like many other mountains within the Rhodope massif, Pirin has a complex geological structure. Three granitic bodies (intrusions) build up the core of the massif: one in the farthest north, one in the center, and one in the south. Around, and between them, the mantle of metamorphic rocks includes

gneisses, schists, amphibolites and marble. Sedimentary rocks are exposed in the periphery, which is surrounded from all sides by faults. The mountain represents a classical horst morphostructure. Volcanic rocks occupy the far east of the area. They are a result of eruptive activities along the Mesta graben during the Paleogene.

In Pirin, marble is found in two main zones: in Northern and in Middle Pirin. To the north, this is the highest section of the main ridge from Vihren peak on the south, to Okaden peak on the north, with the adjacent northeast slope (on the SW slope marble occupies only the highest altitudes). This is a high mountain section about 5 km long, which includes the highest peaks in Pirin: Vihren (2914 m a.s.l.), Kutelo (2908 m a.s.l.), Banski suhodol (2884 m a.s.l.) and Bayuvi dupki (2856 m a.s.l.). Five vast cirques were carved on the northern slope, which overlap over preglacial forms – in fact these are complex glaciokarstic depressions (cirques–uvalas, as mentioned for Pirin by Popov, 1962; and for Durmitor Mountains by Djurović, 2011). From SE to NW these are: Kazanite (Goliam Kazan and Malak Kazan), Kutelo, Banski suhodol, Bayuvi dupki, Kamenitica. This area is the only in Bulgaria where glaciokarst has been developed (Gachev, 2017). In Northern Pirin, marbles occupy also a small spot around Sinanitsa peak to the SW of Vihren.

The marble area is adjacent to silicate metamorphic rocks to the northwest, and with granite to the southeast, at altitudes up to 2600 m a.s.l., are favorable for the existence of fossil rock glaciers.

In middle Pirin marble is spread around Orelek peak (2099 m a.s.l.). Due to the lower altitudes, periglacial landforms are less common, and there are no rock glaciers found in that area.

The high part of Pirin was subjected to extensive glaciation during the Pleistocene. The equilibrium line altitude (ELA) during the maximum glacial stage (not dated) was at 2200–2300 m a.s.l. (Georgiev, 1991). Glaciers were of a valley type, descending 7–11 km to the foot of the mountain at 1100 m a.s.l., along the north heading valleys of Banderitsa and Demianitsa, which join shortly above the town of Bansko (Lilienberg and Popov, 1966). The area above 2200 m a.s.l. is dissected into several large and a number of smaller cirques. The most exten-

sive are those in the central (granitic) part of Pirin: Banderitsa cirque, the cirques of Valiavishki, Prevalski and Vasilashki lakes, Belemeto, Kremenski, Popovo lake cirque and many others (Mironski et al., 1970). Within this area there are around 110 glacial lakes which are scattered across cirque floors (Valkanov et al., 1964).

Apart from two tiny small glaciers (glacierets) in the marble cirques of Pirin, with a total area of 1.5 ha, and several ice patches (Gachev et al., 2016), all glacial landforms are a relict character. Nowadays the high mountain areas are subjected to cryogenic (periglacial in broad sense) activity. Frost weathering is especially intensive on rocky slopes, high summits and ridge tops.

Climatically, Pirin belongs to the sub-mediterranean zone (it is 110 km away from the Aegean Sea). The mountain has the highest range of annual air temperatures in Bulgaria, from +14°C at Kresna, to –1.4°C at Vihren peak (Velev, 2010; [www.swu.bg](http://www.swu.bg)). The 0°C annual isotherm is presently situated between 2650 and 2700 m a.s.l. In general, in high cirques (2400 m. a.s.l. and above) the period with negative temperatures lasts from November to April (and is associated with snowfalls), while between May and October average daily temperatures are mostly positive and rainfalls prevail (Grunewald et al., 2016).

The number of days with transition through 0°C ranges between 90 at 2000 m a.s.l. and 80 at 2900 m a.s.l., and the number of frost days (with minimum daily temperature below 0°C) range between 170 and 260 respectively (Grunewald and Scheithauer, 2008, 2011). In result, exposed ridges at altitudes above 2600–2700 m a.s.l. are strongly weathered. On the other hand, the great number of frost and icy days at cirque floors at 2200–2400 m a.s.l. 200–220 days with minimum temperature below 0°C and 110–130 days with only negative temperatures (Grunewald and Scheithauer, 2008) determines the 6–7 months of snow cover duration at these sites, and the restricted frost action there. Although Pirin is among places with highest annual precipitation amounts in Southern Bulgaria: around 1000–1100 mm for the high mountain zone, the overall sums are small compared to other moun-

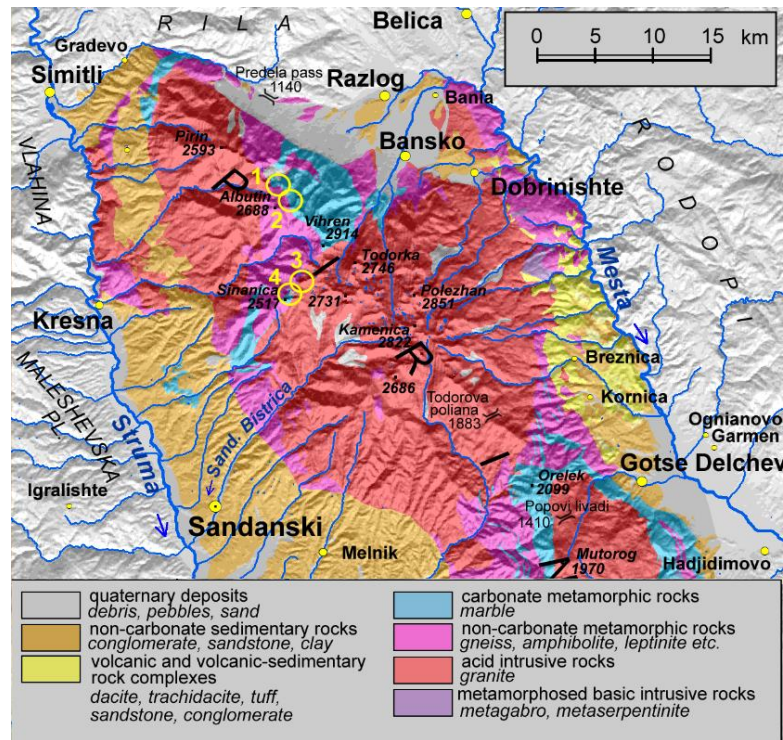
tains in Southern Europe, due to the rain shadow that the Dinaric Mountain and Carpathian mountain ranges produce over of all the eastern part of the Balkans. Around two thirds fall in the cold part of the year (October to March) which provide relatively good snow accumulation.

The mentioned rain shadow effect should have occurred in a similar way in the past, during glacial and late glacial times. Especially in the latter period, the combination of low temperatures and modest precipitation supported the formation of rock glaciers and permafrost. Based on climate reconstructions, it is considered that the rock glaciers in the mountains of Bulgaria (Rila and Pirin) were originally formed during late glacial, in the course of deglaciation (Dimitrov and Gikov, 2011; Dimitrov and Velchev, 2012).

### 3. Methods of research

A detailed geomorphological mapping was done of those cirques in Northern Pirin, which lie on the zone of contact between the marble and the surrounding silicate intrusive and metamorphic rocks: Sinanitsa, Georgiitsa, Razlozhki suhodol and Okaden cirques (Fig. 2). Because of high portion of non-ice component, rock glaciers are usually able to retain much of their morphology long after the permafrost body has completely thawed (Hughes et al., 2003). This was used to identify relict rock glaciers on the field.

Topography maps of 1:10,000 scale were used as a base for the creation of a digital model in ArcGIS 10. Landforms were identified on the field and their main morphometric characteristics were measured with a laser range finder (900 m range). Lithology was diagnosed using geological maps and visually, on the basis of color, crystal structure and lichen cover. The state of lichens was used to make a relative dating of geomorphic activity of rock glaciers and screes. In situ climatic measurements were organized at several sites in the area: on Vihren peak –air temperature and humidity has been measured since October 2014.



**Figure 2** Geological map of Pirin, with the target cirques: 1 – Okaden, 2 – Razlozhki suhodol, 3 – Georgiitsa, 4 – Sinanitsa

#### 4. Representative rock glaciers in mixed lithology

Our observations in the marble (glaciokarstic) high mountain area of Northern Pirin have shown that no rock glaciers have been formed in that area, despite the high altitude and the optimal climatic conditions that might have existed in the past. Only two rock glaciers have been described and studied in cirques where marble contacts with other silicate rocks. They are located in the cirques Razlozhki suhodol and Sinanitsa.

##### 4. 1 Razlozhki suhodol rock glacier

The cirque of Razlozhki suhodol lies on the geological contact between marble (to the east) and gneiss (to the west) (Fig. 3 and 4). The cirque has a NNW aspect. The length from the main ridge to the lowest rim is 3.9 km. the uppermost section of the cirque is approximately 1.3 x 1.3 km, with an area of 2.78 km<sup>2</sup> (Mitkov, 2020). The highest point of that part of the cirque (Razlozhki suhodol peak) is at 2660 m a.s.l., and the lowest altitude is 2320 m a.s.l. The eastern slope of the cirque is higher, in fact a giant rock face of marble, which to the north de-

scends with several toothed peaks, known as 'The stairs'. A line of large screes runs along the slope foot. On the contrary, the western slope, carved in gneiss, is steadier, with a much lower altitude. The gneiss part of the cirque floor is filled with block material, which compose a chaotic pattern of ridges and furrows – a relict rock glacier. The eastern flange of the rock glacier (more than half) is made of marble blocks, which have originated from the marble rocky peak to the south. Descending north, the debris enter over the silicate terrain, where in fact the body of the rock glacier is (Figs. 3 and 4). The existence of the rock glacier in its shape is due to that silicate bed, which at the time of the rock glacier formation, hindered the escape of ground waters and allowed for the formation of permafrost.

The western, much smaller flank of the rock glacier, is entirely made of gneiss blocks. The rock glacier is situated at altitudes between 2410 and 2340 m a.s.l., and occupies an area of about 7.5 ha. In the lower end, debris accumulations are dammed by a ridge of deposits of mixed lithology and size with a steep front – probably a glacial moraine. A small pond (Suhodolsko Lake) is formed before the ridge.



**Figure 3** Razlozhki suhodol cirque: a look towards south

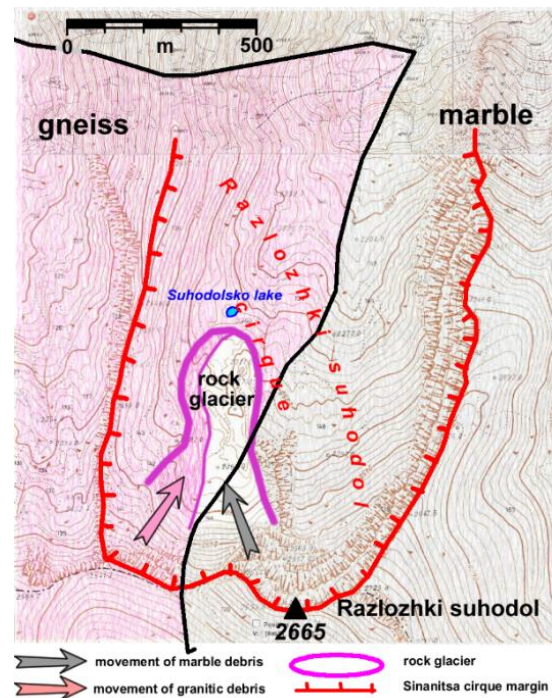
#### 4. 2 Sinanitsa rock glacier

Sinanitsa rock glacier is located in the cirque to the north of the 2517 m high Sinanitsa peak. The cirque is situated in Northern Pirin, on a side ridge descending to the southwest of the main crest of the mountain. The cirque has a NW aspect, and a staircase vertical cross section. The highest level reaches down to 2179 m a.s.l. at Sinanitsa Lake (Figs. 6 and 7). The lake itself has a length of about 140 m, width 90 m and 11 m depth.

In the area, marbles build up only the massif of Sinanitsa peak. They form “a marble cap”, which in this area is a sole remnant of the metamorphic mantle that once covered the granitic intrusion. The marble rock strata, which represent remains from the southern hip of the Sinanitsa anticline (Zagorchev et al. 2017) descend to southwest, forming a monocline structure. The very small thickness of the remaining marble cover (up to 200–250 m) is illustrated by the fact that in the base of the rock walls of Sinanitsa peak, rock start to change, and transition layers are observed. In fact all the cirque floor is on a silicate bedrock.

The rock glacier is developed on the NE, steepest slope of Sinanitsa peak (Fig. 6). It represents a large field full of scattered marble debris and blocks,

which have produced the familiar pattern of ridges and furrows, which is normally addressed as a rock glacier. The altitude of the rock glacier is 2250–2180 m. a.s.l., and the total area is about 9 ha.



**Figure 4** Razlozhki suhodol cirque. Topographic and geological setting of Razlozhki suhodol rock glacier



**Figure 5** The rock glacier in Razlozhki suhodol cirque



**Figure 6** Sinanitsa peak, with the rock glacier in its NE foot and Sinanitsa lake in the background

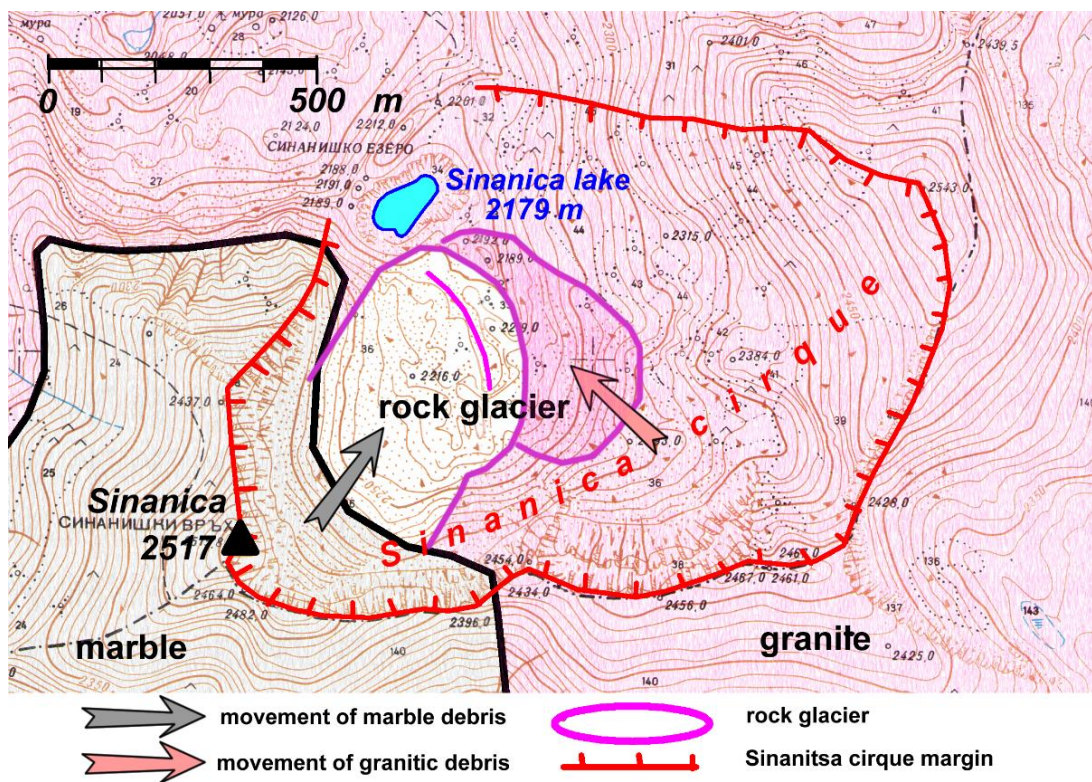
It seems that the NE slope of the peak provided best conditions for formation of abundant debris material – exposed shady rock walls, crossed by numerous tectonic dislocations. In normal situation it is difficult for atmospheric waters to collect in the marble block accumulations, because of the karstic features and the infiltration of waters in depth. But in this case the marble block materials from the slope of Sinanitsa peak have been deposited over a bed of silicate rocks, which allowed for preservation of accumulating of atmospheric waters and for a formation of an active rock glacier in the past, when climate conditions were appropriate.

Like the other of the studied rock glaciers, Sinanitsa rock glacier also has part of it made of blocks

of silicate rock (here it is granite). Debris material for this eastern rock glacier section, are originated from the granitic part of the cirque (see Fig. 7).

On the field, the two lithologically contrasting sections of the rock glacier, which are adjacent to each other, are easily recognizable by the differences in rock and lichen color (Fig. 8): white for marble (down right), greenish for granite (up left).

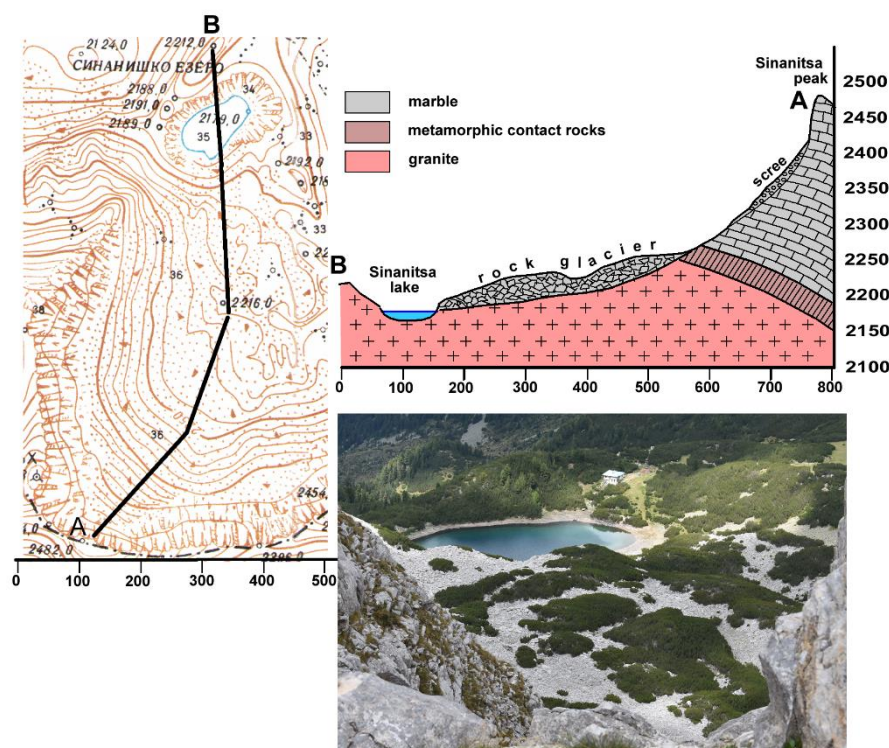
As it was mentioned above, the marble cap of Sinanitsa is quite thin and all the cirque floor is made of granite (Fig. 9). This has allowed for the formation of the rock glacier, and the permanent water body represented by Sinanitsa lake in the northern end of the cirque.



**Figure 7** Sinanitsa cirque, with the location of the relict rock glacier



**Figure 8** A view over Sinanitsa rock glacier: the marble section (in the foreground) is easily distinguished from the granite section (in the background) by the difference in rock coloring



**Figure 9** A profile (cross section) of Sinanitsa cirque and the rock glacier, and a view on the rock glacier from the slopes of Sinanitsa peak

## 5. Formation of rock glaciers in Pirin

According to Dimitrov and Gikov (2011) there are at least 55 rock glaciers in the high mountain zone of the Pirin Mountains, and after more recent studies their number has been increased to 83, at altitudes 2096 to 2710 m a.s.l. (Magori et al., 2017). Most of them are debris rock glaciers, but there are some of talus type (Gachev et al. 2017). Rock glaciers in Rila and Pirin Mountains are mostly considered relict at present (Gikov and Dimitrov, 2010; Dimitrov and Gikov, 2011). It is suggested that they were formed in the period soon after the last retreat of Pleistocene glaciers, i. e. at late glacial time, when the climate was much colder than present, but already unfavorable for classical glaciation (probably in the period Oldest Dryas – Younger Dryas, Dimitrov and Gikov, 2011; Dimitrov and Velchev, 2012).

Climate reconstructions (Bozhilova, 1978; Stefanova and Ammann, 2003; Stefanova et al., 2003) suggest that in general the late glacial in the mountains of SW Bulgaria was relatively dry. This enhanced glacier retreat, and by the middle of the period (Oldest Dryas) all glaciers were already in cirque phase (Mitkov, 2020). The cold dry conditions sup-

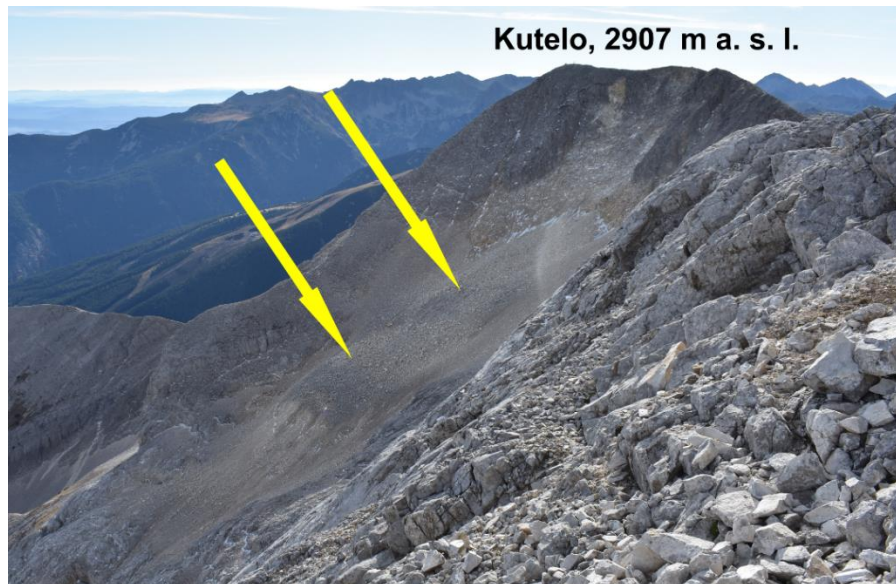
ported the enhanced frost weathering in the high mountain zone. Having in mind the present climate which provides optimal frost heave conditions for the open slopes and ridges at 2500–2900 m a.s.l., it would be suggested that during long time throughout the late glacial, similar conditions prevailed at altitudes 2100–2500 m, where most of Pirin rock glaciers are situated (including the two cases of the present work).

As it was already mentioned, almost all rock types in the high mountain zone of Pirin have potential to produce debris material suitable for rock glacier formation. Apart from the appropriate climatic conditions, which prevailed in the late glacial, proper topographic and geological settings were also crucial for the development of these landforms. Rock glaciers formed mainly at sites where shady or partly shady aspects combine with zones of weakened rocks. The latter are associated with tectonic disruptions of various orientation and size that formed as a consequence of intrusion buildup. In general, the rise of the Pirin granitic intrusions in the course of the Cenozoic caused the gradual folding of the metamorphic series, formed anticline struc-

tures, among which is the Sinanitsa anticline (Zagorchev et al. 2017). Moreover, in the course of multistage intrusion rise and cooling (Zagorchev 1995, 2001), vertical displacements occurred between major fragments within the granitic bodies. All these tectonic activities produced planes and strips of milonitized rocks which have been more easily eroded. Some weakened zones are associated to the magmatic dykes that cut through the cooling granitic and metamorphic bodies and contained still hot magma from the lower parts of magma chambers (Machev, 1993).

Some of the rock glaciers may have evolved from cirque glaciers, with their gradual burial by block and debris material. Others might have formed during subsequent cooling episodes, in re-

sult of freezing of atmospheric waters in the scree slope deposits. Most rock glaciers are found in the granitic part of Pirin, and a fewer number – in areas made of gneiss. Despite the abundance of produced debris material of a proper size, no rock glaciers have been recorded in the high mountain areas built of marble, the only exception being two lobes on the northern slope of Kutelo peak (Fig. 10), which do not possess classical rock glacier morphology. However, the two rock glaciers described in the present work (in the cirques Razlozhki suhodol and Sinanitsa) can also be considered exceptions – in fact they are the only ones that have classical ridge and furrow morphology, and are constituted by marble blocks and debris. This is due to their water impermeable silicate bed.



**Figure 10** The debris lobes in marble, on the northern slope of Kutelo peak – the only landform in the glaciokarstic zone of Pirin which might have been formed due to former permafrost action

In the course of climatic warming throughout the Holocene, temperatures in the lower part of the high mountain belt (2000–2500 m a.s.l.) became too high to sustain permafrost, and all rock glaciers at such altitudes presently have a relict appearance – they are covered in lichens and are vegetated by grasses, junipers and dwarf (mugo) pine. No rock glaciers were formed during the Little Ice Age, although some minor displacements could have occurred at the highest rock glaciers (near the peaks Musala, Goliam Kupa in Rila, Polezhan and Bunferishki chukar in Pirin). Some sporadic permafrost is

supposed to still exist at these sites even at present (Magori et al. 2017; Onaca et al. 2020).

## 6. Conclusions

Pirin is the mountain with the greatest number of rock glaciers in the high mountain zone of Bulgaria. These typical cryogenic landforms are suggested to have formed in late glacial time, in the course of the gradual retreat of Pleistocene glaciers in the highest parts of the Pirin Mountains. Out of over 80 rock glaciers in the massif, only two have been found and described in the highest mountain ridge of Pi-

rin, which is made of marble, as in general marble appears as a rock type which does not support rock glacier formation. In result of this study it has been found that those two rock glaciers were formed in result of favorable accidental combination of topographic and lithological factors, in cirques where the marble contacts with other silicate rocks. In both researched cases (in Razlozhki suhodol and Sinanitsa cirques) marble builds of the highest slopes, which have served as debris source zone for the rock glaciers; but the cirque bed further down is made of silicate rocks. In this way, due to lack of karstic caverns, at the time of rock glacier formation, atmospheric waters that entered the blocky marble material were prevented from escaping down, and were able to remain in the deposits and freeze there, forming the ice core necessary for rock glaciers to move.

In other words, on both sites, rock glaciers had the mechanism of formation typical for silicate geological environments, but with blocky and debris substrate made of marble. Nowadays, when the studied rock glaciers are no more active, the settled marble material is subjected to surface karstification.

## Acknowledgements

This research was supported by South–West University “Neofit Rilski” through the project RP A4/18: Research of high mountain climate through cryospheric indicators: glaciers and permafrost – part 2 (2018), and project RP A 3/20. The high mountains on the Balkans in the conditions of climate change”.

## References

- Barsch D. 1988. Rockglaciers. In: Clark, M. J. (ed.) *Advances in periglacial geomorphology*. John Wiley and sons, New York, 69–87.
- Barsch D. 1996. *Rockglaciers: Indicators for the Present and Former Geoecology in High Mountain Environments*. Springer–Verlag Berlin Heidelberg, Berlin, Germany 331 pp.
- Berthling I. 2011. Beyond confusion: Rock glaciers as cryo-conditioned landforms, *Geomorphology*, **131**: 98–106, /doi.org/10.1016/j.geomorph.2011.05.002, 2011.
- Bozhilova E. 1978. Vegetation changes in the Rila Mountains in the last twelve thousand years. *Annual Journal of Sofia University, Fac. Biol., book–2, Botany*, **71**: 39–44.
- Capps SR. 1910. Rock glaciers in Alaska. *Journal of Geology*, **18**: 359–375.
- Dimitrov P., Gikov A. 2011. Identification and mapping of relict rock glaciers in Pirin Mountains using aerial and satellite images. *Proceedings of the 7th scientific conference “Space, Ecology, Safety”, ISRT, BAS*. (Bg).
- Dimitrov P., Velchev A. 2012. The relict rock glaciers as a morphological feature in Rila mountain’s alpine zone. *Annual Journal of Sofia University, Faculty of Geology and Geography, book 2, Geography*, vol. **103**: 97–112.
- Djurović, P. 2011. High mountain karst on Durmitor. Belgrade, 180 p.
- Gachev E. 2017. High mountain relief in marble in Pirin Mountains: structure, specifics and evolution. *Revista de Geomorfologie*, **19**: 118–135 DOI 10.21094/rg.2017.012.
- Gachev E., Stoyanov K., Gikov A. 2016. Small glaciers on the Balkan Peninsula: state and changes in the last several years. *Quaternary International*, **415**: 33–54.
- Georgiev M. 1991. *Physical Geography of Bulgaria*. Sofia University Press.
- Giardino JR., Vitek JD. 1988. The significance of rock glaciers in the glacial–periglacial landscape continuum. *J. Quat. Sci.* **3**(1): 97–103.
- Gikov A., Dimitrov D. 2010. Identification and mapping of the relict rock glaciers in the Rila Mountain using aerial and satellite images. Conference: Space, Ecology, Safety, Sofia, Bulgaria. 252–259. (Bg).
- Grunewald K. Scheitchauer J. 2008. *Klima– und Landschaftsgeschichte Sudosteuropas. Rekonstruktion anhand von Geoarchiven im Piringebirge (Bulgarien)*. Rhombos, Berlin. 180 pp.
- Grunewald K. Scheitchauer J. 2010. Europe’s southernmost glaciers: response and adaptation to climate change. *Glaciology*, **56**: 129–142.
- Grunewald K. Scheitchauer J. 2011. *Landscape development and climate change in Southwest Bulgaria (the Pirin mountains)*. Springer. 162 pp.
- Grunewald K, Gachev E., Kast G, Nojarov P, Panayotov M. 2016. *Meteorological observations in National park Pirin*. Bansko, Sofia, Dresden. 24 p.
- Haeberli W. 1985. Creep of mountain permafrost: internal structure and flow of alpine rock glaciers. *Mitteilungen der Versuchsanstalt für Wasserbau, Hydrologie und Glaziologie*, ETH Zurich, **77**.
- Hughes PD., Gibbard PL., Woodward JC. 2003. Relict rock glaciers as indicators of Mediterranean palaeoclimate during the Last Glacial Maximum (Late Würmian) in northwest Greece. *Journal of Quaternary Science*, **18**: 431–440. doi:10.1002/jqs.764.

- Lilienberg D, Popov V. 1966. Noie danni ob oledeneni masiva Pirin (Rodopi). (New data about the glaciation of Pirin massif (Rhodopes)). *Doklady AN SSSR*, **167**(5).
- Machev Ph. 1993. Petrology of the Central Pirin pluton. Review of the Bulgarian Geological Society. LIV **1**: 123–137.
- Magori B., Onaca A., Gachev E., Urdea P. 2017. The geomorphological characteristics of rock glaciers and protalus ramparts in the Rila and Pirin Mountains. *Proceedings of the 9th International Conference on Geomorphology*, New Delhi, abstract 302.
- Martin E., Whalley B. 1987. Rock Glaciers Part 1: Rock glacier morphology. *Progress in Physical Geography*, **4**(2): 260–282.
- Mironski N., Valchev A., Popov V. 1970. *Pirin. A guide-book*. Meditsina i fizkultura, Sofia, 182 p.
- Mitkov I. 2020. *Evolution and present state of glacial and cryogenic relief in the marble parts of the Pirin Mountains*. PhD thesis, South–West University, Blagoevgrad.
- Nojarov P. 2012. Variations in precipitation amounts, atmosphere circulation, and relative humidity in high mountainous parts of Bulgaria for the period 1947–2008. *Theoretical and Applied Climatology*, **107**(1): 175–187.
- Onaca A., Ardelean F., Ardelean A., Magori B., Sirbu FS., Voiculescu M., Gachev E. 2020. Assessment of permafrost conditions in the highest mountains of the Balkan Peninsula. *Catena*, **185**(2020) 104288.
- Popov, V. 1962. Morphology of Golemia Kazan cirque in Pirin Mountains. Announcements of the Institute of Geography, Bulgarian Academy of Science. **VI**, 1962, 85–100. (Bg).
- Stefanova I., Ammann B. 2003. Lateglacial and Holocene vegetation belts in the Pirin Mountains (Southwestern Bulgaria). *The Holocene*, **13**/1: 97–107.
- Stefanova I., Ognjanova–Rumenova N., Hofmann V., Ammann B., 2003. Late Glacial and Holocene history of the Pirin Mountains (SW Bulgaria): a paleolimnological study of Lake Dalgoto (2310 m). *Journal of Paleolimnology*, **30**: 95–111.
- Valkanov, B. 1964. Lakes in Bulgaria. *Trudove na IHM (Works of the Institute of Hydrology and Meteorology)*, Sofia.
- Velev S. 2010. *The climate of Bulgaria*, Heron press, Sofia.
- Wahrhaftig C., Cox A. 1959. Rock glaciers in the Alaska range. *Bulletin of the geological society of America*. **70**: 383–436.
- Washburn AL. 1979. *Geocryology: a survey of periglacial processes and environments, second edition*. London: Edward Arnold.
- Zagorchev, I. 1995. Pre–Paleogene Alpine structure of Southwest Bulgaria. *Geologica balcanica*, **26**(5–6): 91–112.
- Zagorchev I. 2001. Tectonics of the Pirin horst. International Conference “Rhodope Geodynamic Hazards, Late Alpine tectonics and Neotectonics”. <http://www.geology.bas.bg/rgh/pirin.html>.
- Zagorchev I., Balica C., Kozhoukharova E., Balintoni IC. 2017. Pirin metamorphic and igneous evolution revisited in a geochronological frame based on U–Pb zircon studies. *Geologica balcanica*, **46**(1): 27–63.
- Bulgarian site for discussing weather [www.stringme-teo.com](http://www.stringme-teo.com).
- Meteorological Punct Vihren. South–West University Neofit Rilski. [www.swu.bg](http://www.swu.bg)

---

# Snow–avalanche history reconstructed with tree rings in Parâng Mountains (Southern Carpathians, Romania)

Corina TODEA<sup>1</sup>, Olimpiu Traian POP<sup>1\*</sup>, Daniel GERMAIN<sup>2</sup>

<sup>1</sup>Laboratory of Dendrochronology, Faculty of Geography, Babeş–Bolyai University, Cluj–Napoca, Romania

<sup>2</sup>Département de Géographie, Université du Québec à Montréal, Montréal, Canada

Received 4 December 2020; Revised 16 December 2020; Accepted 17 December 2020

\*Correspondence to: Olimpiu Traian POP; e-mail: [olimpiu.pop@ubbcluj.ro](mailto:olimpiu.pop@ubbcluj.ro)

---

## ABSTRACT

Snow avalanches are a common phenomenon in Parâng Mountains (Southern Carpathians, Romania) perturbing tourism activities and associated infrastructures, damaging forests, and causing fatalities. Its past history is an essential information to gather while assessing the hazard zonation areas. Usually, in Romania snow–avalanche activity occurring in forested areas are neither monitored, nor recorded by historical archives. In these areas, environmental archives such as tree rings may provide useful information about the past avalanche activity. The purpose of the present study is to reconstruct snow–avalanche history with tree rings along a path located below Cârja Peak (2405 m a.s.l.), an area where past snow–avalanche activity still remains underestimated. In this sense, 57 Norway spruce (*Picea abies* (L.) Karst.) trees showing clear signs of disturbance by snow avalanches were sampled and the growth anomalies associated with the mechanical impact produced by snow avalanches on trees were identified within their rings and served to reconstruct past events. The reconstructed chronology covers the period 1994–2018 showing the occurrence of a minimum of 11 major events, with an average return period of 2.1 years. Tree–ring records provided the most consistent avalanche event chronology in the study area. Although the limited extension of the chronology back in time, a better understanding of snow–avalanche history which may be gained through dendrochronological reconstructions represent nonetheless useful and pertinent information to consider before the implementation and development of infrastructure in this mountain avalanche–prone area.

**KEYWORDS** snow avalanches, dendrogeomorphology, return periods, *Picea abies* (L.) Karst., Parâng Mountains (Romania)

---

## 1. Introduction

Snow avalanches (SAs) are known as a common phenomenon occurring in snow–covered mountainous areas worldwide, being one of the main

agents of debris transfer in such environments (e.g., Rapp, 1959; Gardner, 1970; Ward, 1985; Decaulne and Saemundsson, 2006). As they can frequently endanger constructions, affect tourist activities and related infrastructures, and cause fatalities especially

during backcountry recreation, it is of paramount importance to evaluate the risk related to the dynamics of snow avalanches for an accurate hazard zonation, but also to prevent future exposure of people and tourist infrastructures in such avalanche-prone areas. The past SAs history, especially regarding their magnitudes and frequencies are an essential information to gather while assessing the hazard zonation areas. However, SAs activity occurring in remote forested mountain areas are neither monitored, nor recorded by historical archives. In these areas, environmental archives such as tree rings may provide useful information about the past frequency and magnitude of avalanche activity (Schaerer, 1972; Rayback, 1998; Boucher et al., 2003; Germain et al., 2005; Casteller et al., 2007; Reardon et al., 2008; Muntán et al., 2009; Corona et al., 2010a, 2012; Decaulne et al., 2012; Voiculescu and Onaca, 2013; Šilhán and Tichavský, 2017). Tree-ring investigations rely on the assumption that annual growth rings record accurately and objectively, in their anatomical structure, the mechanical impact of snow avalanches on trees.

In the last decades, the western and eastern flanks of the Parâng Mountains (Southern Carpathians, Romania) have become an attractive area for winter recreational activities. This recent increase in tourist numbers, coupled with the projected increase with the development of appropriate infrastructure to sustain winter recreational activities, was planned in an avalanche-prone area, which could result in an increased exposure of people and infrastructure to avalanche hazards. In this regard and considering the lack of knowledge regarding SAs activity in this high mountain area, the absence of forecasting and warning services, tourists may cross either unconsciously or even consciously some avalanche paths and hazard zones. Giving the availability of trees as natural archives on these forested slopes on the SW flank of Parâng Mountains, tree-ring-based snow-avalanche reconstructions have already been established along two adjacent avalanche paths and for several decades, up to one century (Pop et al., 2016; Meseșan et al., 2019). However, these two avalanche paths showed discrepancies in avalanche regime, likely related to their slope aspect as well as their morpho-

metric characteristics. Nearby, another avalanche path appears of interest considering the tragic event that occurred on December 1<sup>st</sup> 2017, when an avalanche released below Cârja Peak (2405 m a.s.l.) and buried two people, one of whom later died. SAs activity along the path below Cârja Peak still remains underestimated, but a dendrogeomorphic investigation does provide useful information for a better knowledge of the avalanche regime, and subsequently adequate zoning. The aim of the present study is therefore to reconstruct snow-avalanche history with tree rings along the path below Cârja Peak in the Parâng Mountains. From the hazard management perspective, a better understanding of SA history in the area which may be gained through dendrochronological reconstructions should contribute significantly to hazard zoning and the establishment of appropriate risk mitigation measures.

## 2. Study area

The Parâng Mountains belongs to the Southern Carpathians Range (Fig. 1) and are one of the highest Mountain Range in Romania, with crests and peaks above 2500 m (Parângul Mare Peak, 2519 m a.s.l.). The complex geology of the area is characterized by schists and granitic intrusions in the central mountain area, and sedimentary rocks amongst them with limestones in the southern part (Badea et al., 2001). Evidence of Pleistocene and Holocene glacial and periglacial activity is well-attested by the alpine landscape morphology: narrow crests, simple and complex glacial cirques, nivation cirques, U-shaped glacial valleys, moraine deposits, etc. (Urdea et al., 2011). The glacial landforms are better developed on the northern slopes. On steep slopes above 25°, SAs are a widespread present-day process and particularly at high-elevation with glacial and periglacial ones (Gavrilă et al., 2018). The timberline is found between 1700 and 1800 m a.s.l., but avalanches are lower than this limit as evidenced by the presence of corridors devoid of mature forest vegetation along upper reaches valley. The avalanche path investigated (45°22' N, 23°35' E) is located on the western flank of Parâng Mts., below Cârja Peak (2405 m), ranging between 2200 m and 1500 m a.s.l. (Fig. 2). The avalanche path as delineated on the orthophotos has a surface area of 910.466 m<sup>2</sup>

(based on a DEM with a 10-m accuracy). The slope profile shows steep slopes (25° to 45°) in the starting and the track zones and lower slope angles (< 25°) in the runout zone. The morphology of the valley limits the lateral spread of avalanches, resulting then in a typical funnel-shaped avalanche path. The path is clearly visible by comparison to the adjacent forest. Along the track and runout zones, one can observe that tree cover is replaced by a mix of

shrubs (*Pinus mugo* and *Juniperus communis*) and sparse coniferous (*Picea abies* (L.) Karst.) and deciduous (*Acer pseudoplatanus*) trees related to the frequency of avalanches. In the starting zone, the path is crossed by the main hiking trail used by tourists that follow the mountain crest, which represents a significant risk, especially during winter and spring seasons. Lower in the track zone, the path is also crossed by an additional unmarked hiking trail.

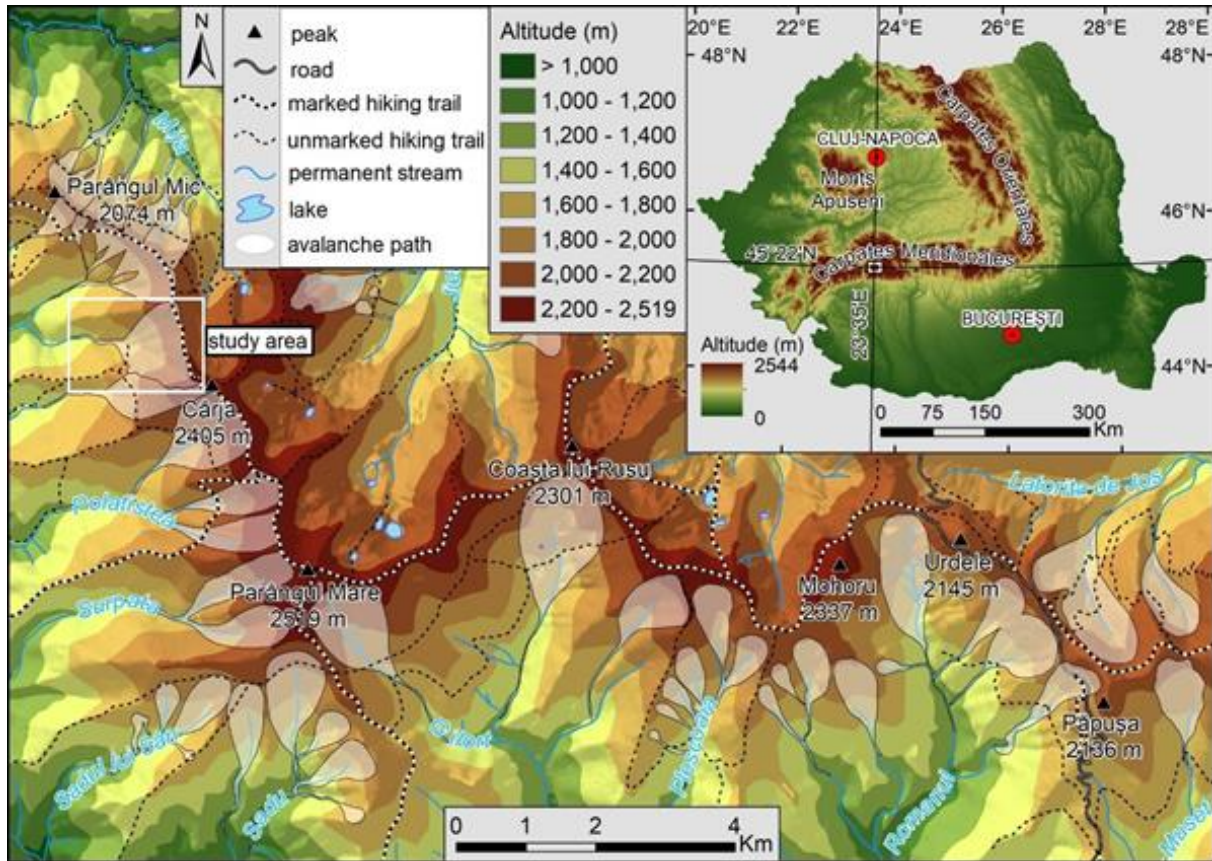


Figure 1 Location of the study area

### 3. Materials and methods

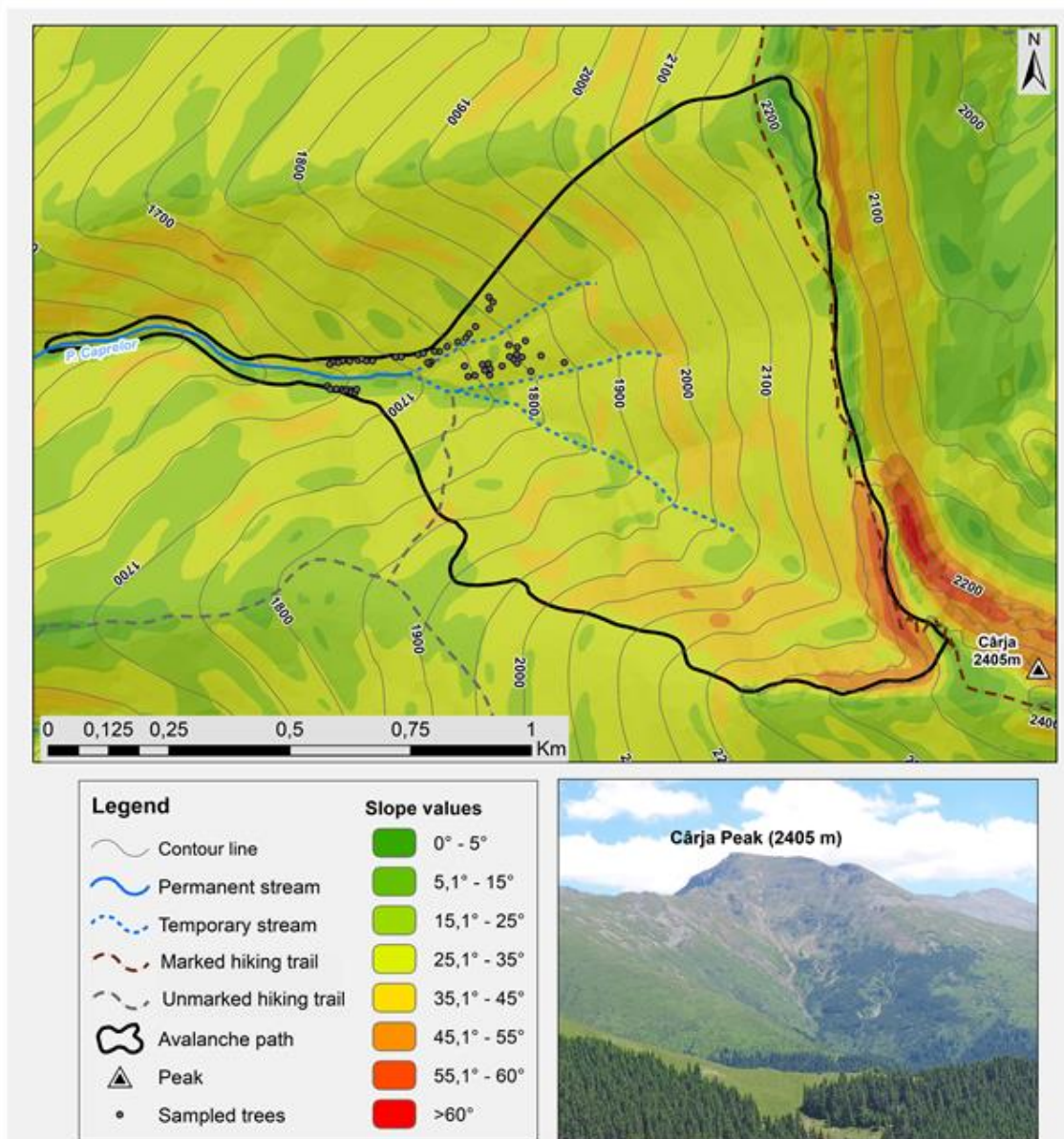
#### 3.1 Field sampling

The path investigated was mapped using a 1:25000 scale topographical map and 1:5000 scale orthophotoplans (2012 campaign). A fieldwork campaign was conducted in August 2018. During fieldwork, a total number of 57 Norway spruce (*Picea abies* (L.) Karst.) trees were sampled. The samples were collected only from trees showing visible disturbances caused by the mechanical impact of avalanches, e.g., tilted stems, injured stems, decapitated trees, trimmed crown, and partially uprooted trees. Sam-

ple collection included either discs cut from 29 trees, as well as cores extracted from 28 trees (2, 3 or 4 cores per tree, 60 cores in total). Even not very large, the sample depth is coherent with the spatial extent of the path in the study site, as well as with the tree-cover conditions. The sampling strategy consisted also to collect samples from trees belonging to different age classes. Younger trees with high stem flexibility are more sensitive to the recent avalanche activity, showing then stronger reactions in their annual rings. By comparison, older trees are considered on one hand less sensitive to recent ac-

tivity, but they can provide on the other hand a longer SAs chronology as they might have recorded events in the earlier decades of their lifespan (Stoffel et al., 2013; Stoffel and Corona, 2014). For each sampled tree, additional information was gathered, including tree height, the social position of the tree within the forest, nature and position of visible disturbance on stems, the position and height of the samples on the stem, as suggested by Stoffel and Bollschweiler (2008). Besides the information regarding tree morphological characteristics, photos of each sampled tree were gathered, as well as their

location by using a GPS device. In addition, 22 spruce trees located outside the avalanche path in a forest stand undisturbed were also sampled (two cores per tree, 44 cores in total), in order to obtain a local reference chronology, which represents the common growth variation of trees in the area (Cook and Kairiukstis, 1990; Schweingruber, 1996). This reference chronology was used for cross-dating the growth series of disturbed trees and to distinguish the growth anomalies related to mechanical impact of SAs from those related to local stand growth conditions (Stoffel et al., 2006).



**Figure 2** Delineation of the avalanche path and location of sampled trees within the path

### 3.2 Sampling preparation and analysis

The samples were prepared for analysis following the procedures described by Bräker (2002) and Stoffel and Bollschweiler (2008). Working procedure includes core mounting in wood supports, air-drying and sanding of samples using a sanding belt machine. Successive sanding operations were applied using an increasing grits of the sanding belts (80, 180, 220, 400, 600 grits), in order to obtain wood surfaces suitable to reveal clearly the ring boundaries and their anatomical structure. The age of each sample was determined by ring counting under the stereomicroscope. Tree-ring width was then measured along two, three or four radii for cores and along four radii for cross-sections. Ring-width measurements were performed to the nearest of 0.01 mm accuracy using a LINTAB 5 system composed of Leica DMS 1000 stereomicroscope connected to the measurement table and to the computer with TSAPWin™ Scientific software (Rinntech, 2020). Tree-ring series of trees growing outside the avalanche path were also measured to establish a local reference chronology. The calendar year of each ring in the disturbed tree growth chronologies was determined by cross-dating with the local reference chronology. The quality of the cross-dating was visually evaluated using the graphical function of the TSAP-Win software and statistically analysed using COFECHA software (Holmes, 1983).

### 3.3 Growth anomalies identification and event reconstruction

Ring-width series measured on samples from disturbed trees were visually cross-dated against the reference chronology, in order to (i) detect possible false or missing rings and to correct ring-width series of disturbed trees, and (ii) discriminate between growth anomalies related to SAs activity and the normal growth related to climate conditions (Bryant et al., 1989). Corrected growth chronologies of disturbed trees were analysed in order to identify the specific growth anomalies related to the SAs activity within tree-rings. Typical growth reactions related to mechanical impacts of SAs (Stoffel and Corona, 2014), e.g., impact scars (SC), traumatic resin ducts (TRD), compression wood (CW) in association with the onset sequences of abrupt growth suppression

(GS) were identified within rings of disturbed trees. Growth reactions developed in the first decade of the disturbed trees were disregarded for event reconstruction, as young trees having commonly thin stems are more susceptible to disturbances by various factors other than SAs and therefore represent potentially a source of dating errors (Stoffel and Bollschweiler, 2008; Šilhán and Stoffel, 2015; Tichavský et al., 2017). The avalanche chronology was reconstructed using the Avalanche Activity Index (AAI) – an equation which includes the number of trees showing growth anomalies during a particular year ( $R_t$ ) divided by the total number of sampled trees alive in a specific year ( $A_t$ ) (Shroder, 1978, Butler and Malanson, 1985):

$$AAI = (\sum_{i=1}^n R_t / \sum_{i=1}^n A_t) \times 100 \quad (1)$$

To distinguish between the tree growth reactions caused by mechanical impact of SAs from those considered as a noise and recorded by trees, five growth disturbance (GD) intensity classes were defined, as proposed by Stoffel and Corona (2014):

- GD Intensity 5: SC, strong TRD;
- GD Intensity 4: moderate TRD, strong CW, strong GS;
- GD Intensity 3: moderate CW and moderate GS;
- GD Intensity 2: weak CW;
- GD Intensity 1: weak TRD.

The intensity of each reconstructed avalanche event was determined based on the intensity classes (Stoffel and Corona, 2014) of the GDs found within the tree rings. For each class of intensity, a specific value was accorded as reported by Koglenig-Meyer et al. (2011):

$$W_{it} = [(\sum_{i=1}^n T_5 \times 5) + (\sum_{i=1}^n T_4 \times 4) + (\sum_{i=1}^n T_3 \times 3)] \times \frac{\sum_{i=1}^n R_t}{\sum_{i=1}^n A_t} \quad (2)$$

where  $T_5$  is the number of trees with GD intensity class 5, multiplied by a factor 5;  $T_4$  is the number of trees with GD intensity class 4, multiplied by a factor 4;  $T_3$  is the number of trees with GD intensity class 3, multiplied by a factor 3;  $R$  is the number of trees showing GDs in the year ( $t$ ) and  $A$  is the total number of sampled trees alive in the year ( $t$ ).

Finally, a specific year was considered as being an event year if the following three conditions are fulfilled simultaneously: (i) >10 available trees of 10

years old for the reconstruction; (ii) a minimum of three trees showing severe or moderate growth disturbances; and (iii) a minimum AAI of 10%.

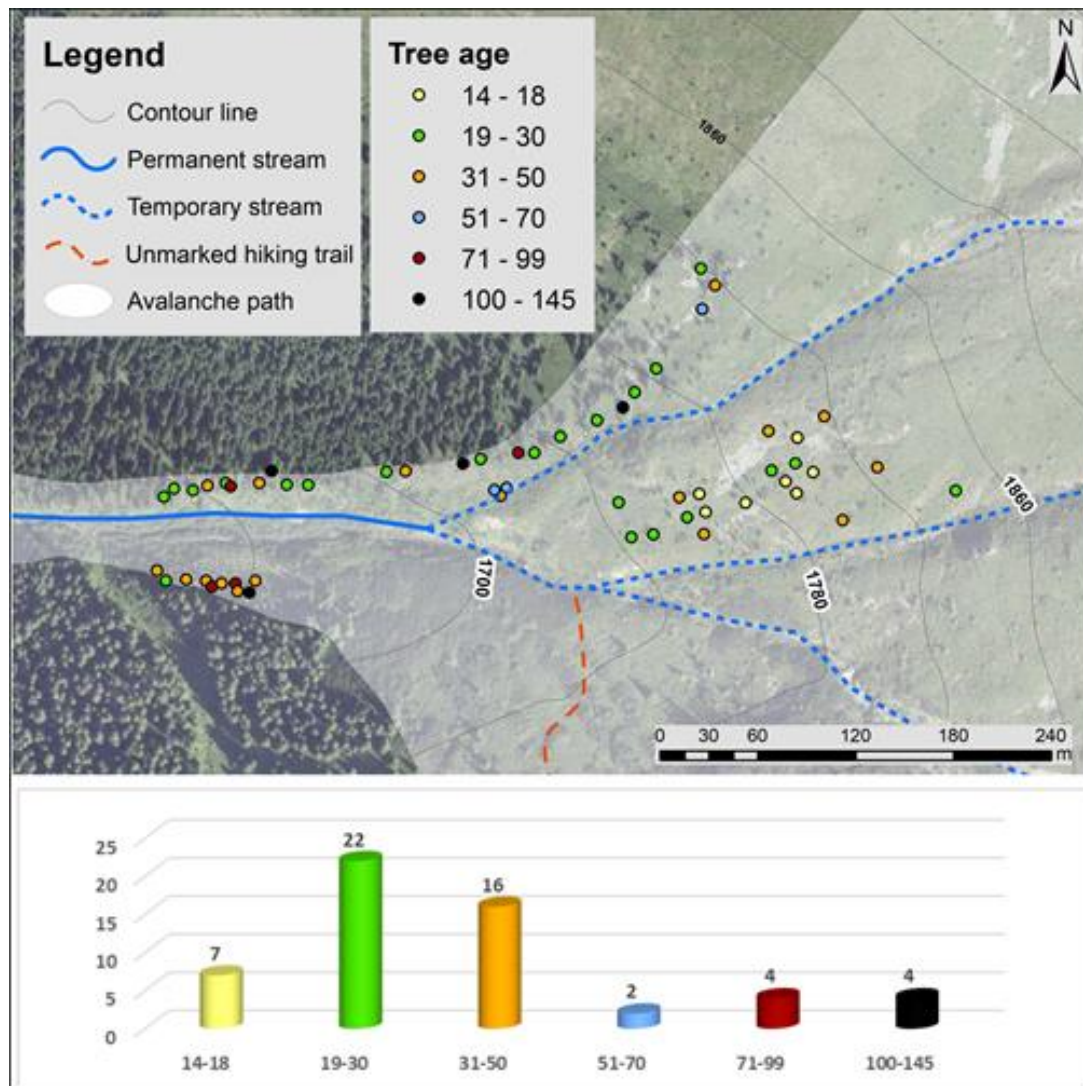
Considering that SAs occur during the winter and spring seasons, which is the dormant period of trees, they can only be recorded in the growth rings of the following summer and fall seasons. For example, the reconstructed event year 2018 means that a SA occurred during the cold season 2017–2018 (winter and spring).

## 4. Results

### 4.1 Age structure of the forest stand

The age of the sampled trees is variable, the youngest tree is 15 years old and the oldest one reaches

145 years old. Figure 3 shows the spatial position of sampled trees and the tree-age classes. One can observe that the older trees are located to the path margins where could indicate less impact of the SAs. A group of young trees (classes 14–18 years old and 19–30 years old) and medium aged trees (class 31–50 years old) is located in the axial area of the path and seems to indicate that the avalanche path affected repeatedly this area and consequently disturbed trees cannot reach significant heights. The right side of the path is characterized by younger trees (class 19–30 year old) than on the left side (31–50 years old). This suggests that SAs flow closer to the right side of the path, having then a higher impact on the forest.



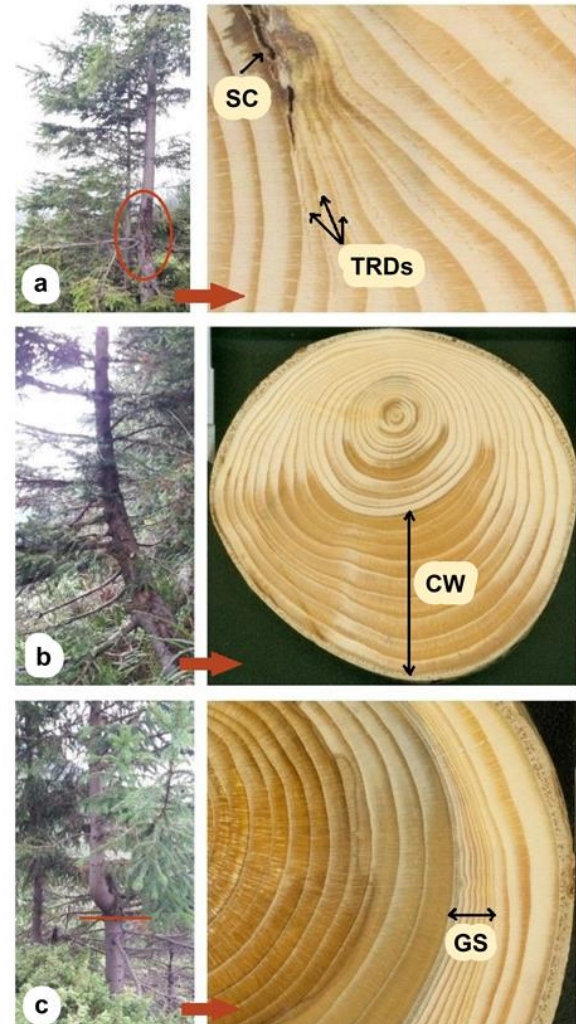
**Figure 3** Spatial distribution and age classes of the sampled trees

## 4.2 Reconstructed avalanche events, frequency and spatial distribution

Tree-rings analyses enabled to identify a total number of 354 growth anomalies, among which 185 growth reactions of severe and moderate intensities (GD intensity classes 3 to 5). Taking into consideration the three criteria above-mentioned in the method section, a minimum of 11 SA events was reconstructed. The chronology spans the period 1994 – 2018. Other possible event years could be reconstructed, but the small number of growth reactions, in combination with a reduced sample depth and an  $AAI < 10\%$  do not enabled us to consider with confidence these years.

The large number of cross-sections available in the sample collection enabled to identify a large number of scars (SCs) and strong traumatic resin ducts (TRDs), two reliable indicators representing higher percentage than all the other types of growth reactions. Therefore, even though the reconstructed chronology is short, the event years that were recorded show strong and undoubtedly signals of snow avalanche activity. Table 1 shows the types and the number of growth disturbances for each reconstructed event year. Compression wood (CW) sequences resulted as a reaction to the stem tilting were also identified within the samples. Series of growth suppressed rings following the tree topping represent a valuable indicator of SAs and can be found as long sequences especially in the case of the event years with high number of trees disturbed (e.g., 2005 and 2007). Figure 4 provides some examples of typical growth reactions identified within the samples and used for event reconstructions. Figure 5 illustrates the event years within the reconstructed chronology which covers a short period of 24 years. The length of the reconstructed chronology is limited by the age of the sampled trees. However, this may also suggest that a recurrent SAs activity destroyed repeatedly the forest cover within the path. The highest number of disturbed trees was recorded in 2005 and 2007, when the percentage of affected trees is above 30% from the total sampled trees. That represents a strong signal of disturbances occurring during the previous dormant seasons, i.e., snow avalanches occurring during winter seasons 2004–2005 and 2006–2007. During winter

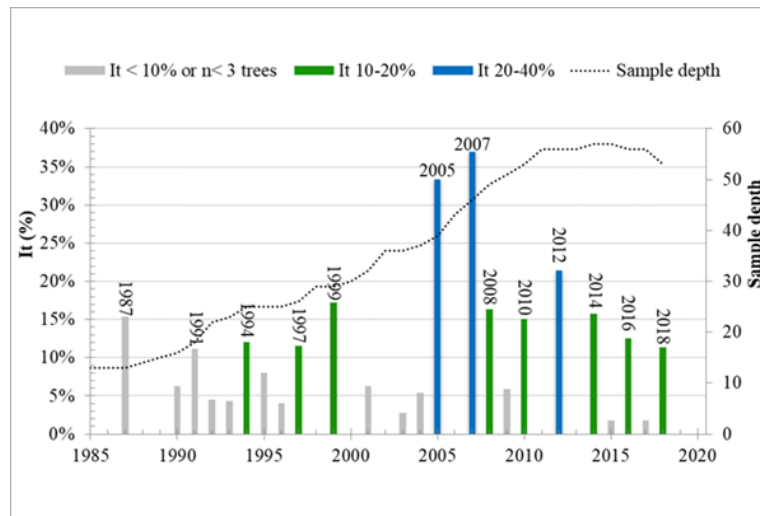
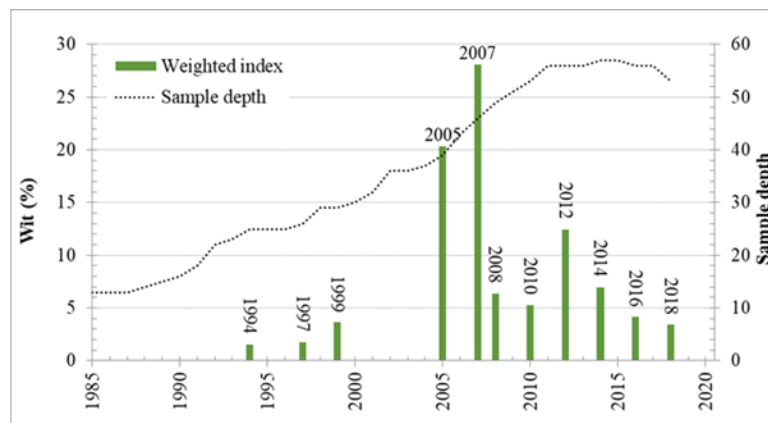
season 2011–2012, several trees were also strongly impacted by SAs, as this year recorded the most scar-type reactions (seven in total) from all event years in the chronology. Few other years (e.g., 1997, 1991) reached the threshold of  $AAI > 10\%$ , but they were not considered as event years because less than three trees showed severe or moderate growth disturbances.



**Figure 4** Examples of disturbed trees sampled and their growth reactions related to mechanical impact of SAs: (a) wounded tree due to mechanical impact of the avalanche and scar (SC) overgrown by subsequent tree rings and associated rows of traumatic resin ducts (TRDs) in the following rings after the disturbance; (b) tilted (sabre-like form) stem and its associated growth reaction, a compression wood (CW) sequence formed in several rings after the disturbance; (c) topped (apex loose) tree stem and associated growth suppressed (GS) rings

**Table 1** Type, intensities and number of growth disturbances identified within rings and used in the SAs reconstruction

|      | SC | strong TRD | moderate TRD | weak TRD | strong GS | moderate GS | weak GS | strong CW | moderate CW | weak CW | Trees aged > 10 | Affected trees | AAI (%) | W <sub>it</sub> (%) |
|------|----|------------|--------------|----------|-----------|-------------|---------|-----------|-------------|---------|-----------------|----------------|---------|---------------------|
| 2018 | 6  | 5          | 0            | 0        | 0         | 0           | 0       | 0         | 0           | 0       | 53              | 6              | 11      | 3.3                 |
| 2016 | 4  | 5          | 2            | 7        | 0         | 0           | 0       | 0         | 0           | 0       | 56              | 7              | 13      | 4.1                 |
| 2014 | 6  | 6          | 1            | 12       | 0         | 1           | 3       | 0         | 0           | 2       | 57              | 9              | 16      | 6.9                 |
| 2012 | 7  | 8          | 4            | 3        | 0         | 0           | 1       | 1         | 0           | 4       | 56              | 12             | 21      | 12.4                |
| 2010 | 3  | 1          | 4            | 8        | 0         | 2           | 1       | 0         | 1           | 0       | 53              | 8              | 15      | 5.2                 |
| 2008 | 5  | 5          | 0            | 3        | 3         | 1           | 1       | 0         | 0           | 1       | 49              | 8              | 16      | 6.3                 |
| 2007 | 5  | 7          | 6            | 3        | 3         | 3           | 5       | 7         | 0           | 3       | 46              | 17             | 37      | 28.1                |
| 2005 | 5  | 8          | 3            | 8        | 4         | 0           | 1       | 1         | 1           | 8       | 39              | 13             | 33      | 20.3                |
| 1999 | 0  | 1          | 4            | 5        | 1         | 0           | 3       | 0         | 0           | 3       | 29              | 5              | 17      | 3.6                 |
| 1997 | 3  | 1          | 0            | 1        | 0         | 0           | 4       | 0         | 0           | 0       | 26              | 3              | 12      | 1.7                 |
| 1994 | 0  | 1          | 2            | 2        | 0         | 0           | 0       | 0         | 0           | 0       | 25              | 3              | 12      | 1.5                 |

**Figure 5** Reconstructed SAs chronology**Figure 6** Intensity of growth disturbances expressed as Weighted index (W<sub>it</sub>) calculated for the reconstructed SAs years

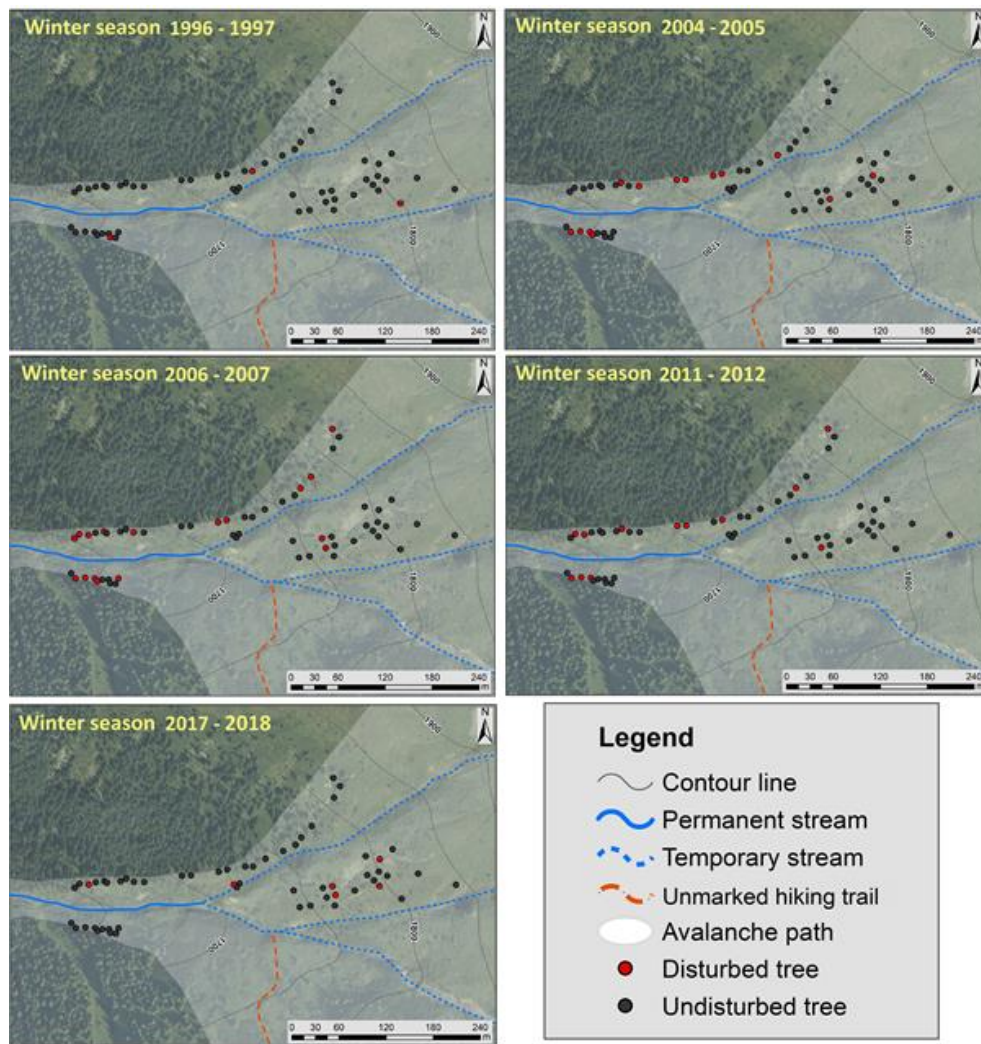
For each event year in the reconstructed chronology, a Weighted Index ( $W_{it}$ ) was calculated based on the intensity of the GDs, in order to assess the impact of the SAs on the sampled trees (Fig. 6). SAs which occurred during the winter seasons 2004–2005, 2006–2007, and 2011–2012 not only have been confirmed by a highest number of affected trees, but also by the highest number of the GDs of strong intensities. The other reconstructed event years show lower values of the  $W_{it}$ . Figure 7 presents some examples of the spatial extension of SAs where in the winter seasons 2004–2005, 2006–2007 and 2011–2012 the entire path was affected.

## 5. Discussion

One of the limits of the dendrogeomorphic approach applied here to reconstruct SAs history is that it enabled to reconstruct the occurrence of a single SA per winter season, even if more than one SA occurred during the same winter period. Moreover, high-magnitude SAs are usually recorded by a high number of trees distributed spatially across the entire path, while low-magnitude SAs may not reach with sufficient velocity and impact all the trees inside the path, and consequently are not recorded. Spatial location of trees within the path has potentially a major influence on the possibilities to reconstruct SAs in different path sectors.

Trees, depending on their location on the left, median or right sides of the upper track area, may potentially be impacted by both low-magnitude and high-magnitude SAs, while trees located in the runout area are disturbed by those major SAs affecting the entire path. In this respect, considering the location of sampled trees within the Cârja path mainly on the left and right sides of the track area, we are aware that not all SAs have been reconstructed based on tree-ring analyses. Therefore, the results of the present study should be considered only as a minimum chronology of SAs activity along the Cârja path. An improved SAs chronology could be obtained by combining the tree-ring reconstruction with historical archives such as reported elsewhere (Corona et al., 2010b; Tumajer and Treml, 2015). In areas devoid of forest vegetation, other indirect information may be obtained regarding the patterns of SAs activity (SAs type, frequencies), ei-

ther on short-term, as well as on long-term period. During successive campaigns, field observations may concern the geomorphic impact of SAs, and particularly the characteristics of transported debris by SAs, e.g., boulder size and orientation, plunging marks, colluvial cone deposit stratigraphy, etc. (Rapp, 1959; Nyberg, 1988; Decaulne and Sae-mundsson, 2006). Unfortunately, the lack of historical records in the Parâng Mountains and therefore tree-ring reconstruction cannot be completed by this source of information. A single event documented by newspapers and witnessed by members of Mountain Rescue Public Service (MRPS) occurred in December 1<sup>st</sup> 2017, and related with the loss of one human life. Therefore, tree-ring reconstruction may be considered as the method closest to absolute dating of SAs activity that can be used presently in the study area. The scientific community shows different opinions regarding the suitable sample size and the minimum threshold value of AAI to consider an avalanche event (Decaulne et al., 2012). By comparing historical archives with dendrogeomorphic reconstructions, Corona et al. (2012) suggested a sample size of approximately 100 trees in order to obtain a more complete SAs chronology. Some authors reported sample sizes less than 20 trees unreliable for avalanche reconstruction (Butler et al., 1987). The minimum threshold of AAI should be variable depending on the sample size, varying from 10% (Boucher et al., 2003; Dubé et al., 2004; Germain et al., 2005; Reardon et al., 2008) to 40% for the identification of high-magnitude events (Butler and Malanson, 1985; Muntán et al., 2009). However, there is no general agreement on defining a specific sampling size which may be applied in all tree-ring reconstructions of SAs. Although a sampling size around 100 trees could be useful in some mountain areas with large paths (Corona et al., 2010b), in other mountain areas with small paths and/or less dense tree cover may not be applied. Choosing a suitable sampling size to reconstruct SAs in an area is still subjective by the expert judgement and depends on various sampling strategies applied, available trees for sampling, sample type which may be obtained (discs, wedges or cores), various resources involved during field-sampling (time, experts, equipment, etc.).



**Figure 7** Spatial distribution of disturbed trees for winter seasons: 1996–1997, 2004–2005, 2006–2007, 2011–2012 and 2017–2018

In the present study, giving the morphometric features of the investigated path (relatively small size, confined path), the small density of tree cover inside the path, the possibility to collect samples of high quality (discs) due to availability of the uprooted trees left after the 2017 SA event, and the limited time resources dedicated to field investigations, a medium sample size was considered as suitable. A minimum threshold of 10% was used and considered suitable to reconstruct past SA events, but we are aware that a larger sample size would likely enable to extend the chronology and improve the historical reconstruction of past SA events. However, it is also interesting to mention that some of the reconstructed SA winters have also been documented in the Romanian Carpathians. Indeed, the years

2005 and 1997 were also recorded in the previous tree. The scientific community shows different opinions regarding the suitable sample size and the minimum threshold value of AAI to consider an avalanche event (Decaulne et al., 2012). By comparing historical archives with dendrogeomorphic reconstructions, Corona et al. (2012) suggested a sample size of approximately 100 trees in order to obtain a more complete SA chronology. Some authors reported sample sizes less than 20 trees unreliable for avalanche reconstruction (Butler et al., 1987). The minimum threshold of AAI should be variable depending on the sample size, varying from 10% (Boucher et al., 2003; Dubé et al., 2004; Germain et al., 2005; Reardon et al., 2008) to 40% for the identification of high-magnitude events (Butler and Malan-

son, 1985; Muntán et al., 2009). However, there is no general agreement on defining a specific sampling size which may be applied in all tree-ring reconstructions of SAs. Although a sampling size around 100 trees could be useful in some mountain areas with large paths (Corona et al., 2010b), in other mountain areas with small paths and/or less dense tree cover may not be applied. Choosing a suitable sampling size to reconstruct SAs in an area is still subjective by the expert judgement and depends on various sampling strategies applied, available trees for sampling, sample type which may be obtained (discs, wedges or cores), various resources involved during field-sampling (time, experts, equipment, etc.). ring-based chronologies either from Parâng Mountains (Pop et al., 2016; Meseşan et al., 2019) and other mountain ranges in the Southern and Eastern Carpathians (Voiculescu and Onaca, 2013; Chiroiu et al., 2015; Pop et al., 2017; Pop et al., 2020). In that regard, further investigations on other avalanche paths in Parâng Mountains are needed, in order to analyse the weather conditions and patterns that are likely favourable to the triggering of SAs at a regional scale in the Carpathian Range.

## 6. Conclusions

The type, amount and intensities of growth disturbances found within the tree rings enabled the reconstruction of SA history for the period 1994–2018. Within the investigated avalanche path, the tree-ring based chronology shows the occurrence of a minimum of 11 significant SA events, with an average return period of 2.1 years. Tree-ring records provided the most consistent avalanche chronology and may further contribute to an accurate avalanche hazard zonation in the study area. As most of the avalanche paths in the area are crossed by tourist hiking trails, reconstruction of SAs activity over the last decades should be extended to the whole Parâng Mts., in order to improve the risk assessment at a regional scale. The results of the present study and other similar studies can help the Mountain Rescue Team and local administration in increasing the awareness about SAs location, frequency, and magnitude. Finally, although the limited extension of the chronology back in time, this represents nonetheless useful and pertinent information to

consider before the implementation and development of infrastructure in this mountain avalanche-prone area.

## References

- Badea L, Niculescu G, Roată S, Buza M, Sandu M. 2001. *Unitățile de relief ale României – Vol. I: Carpații Meridionali și Munții Banatului*. Ed. Ars Docendi, Bucharest, vol. 1, ISBN 973-0-02290-9 (in Romanian).
- Bräker OU. 2002. Measuring and data processing in tree-ring research – a methodological introduction. *Dendrochronologia* **20**(1–3): 203–216.
- Boucher D, Filion L, Hétu B. 2003. Reconstitution dendrochronologique et fréquence des grosses avalanches de neige dans un couloir subalpin du mont Hog's Back, Gaspésie Centrale (Québec). *Géographie Physique et Quaternaire* **57**: 159–168.
- Bryant, C.L., Butler, D.R. & Vitek, J.D., 1989. A statistical analysis of tree-ring dating in conjunction with snow avalanches: Comparison of on-path versus off-path responses. *Environmental Geology and Water Sciences* **14**(1): 53–59.
- Butler DR, Malanson GP. 1985. A history of high-magnitude snow-avalanches, southern Glacier National Park, Montana, USA. *Mountain Research and Development* **5**: 175–182.
- Butler DR, Walsh SJ, Oelfke JG. 1987. Tree-Ring Analysis and Natural Hazard Chronologies: Minimum Sample Sizes and Index Values. *Professional Geographer* **39**: 41–47.
- Casteller A, Stockli V, Villalba R, Mayer AC. 2007. An evaluation of dendroecological indicators of snow avalanches in the Swiss Alps. *Arctic, Antarctic and Alpine Research* **39**: 218–228.
- Chiroiu P, Stoffel M, Onaca A, Urdea P. 2015. Testing dendrogeomorphic approaches and thresholds to reconstruct snow avalanche activity in the Făgăraș Mountains (Romanian Carpathians). *Quaternary Geochronology* **27**:1–10.
- Cook E, Kairiukstis L. 1990. *Methods of dendrochronology: applications in the environmental sciences*. Kluwer Academic Publishers, Dordrecht, The Netherlands, 394 p.
- Corona C, Rovéra G, Lopez-Saez J, Stoffel M, Perfettini P. 2010a. Spatio-temporal reconstruction of snow avalanche activity using tree rings: Jean Jeanne avalanche talus, Massif de l'Oisans, France. *Catena* **83**: 107–118.
- Corona C, Saez JL, Stoffel M, Bonnefoy M, Richard D, Astrade L, Berger F. 2010b. How much of the real avalanche activity can be captured with tree rings? An

- evaluation of classic dendrogeomorphic approaches and comparison with historical archives. *Cold Regions Science and Technology* **74** (75): 31–42
- Corona C, Saez JL, Stoffel M, Bonnefoy M, Richard D, Astrade L, Berger F. 2012. How much of the real avalanche activity can be captured with tree rings? An evaluation of classic dendrogeomorphic approaches and comparison with historical archives. *Cold Regions Science and Technology* **74–75**: 31–42.
- Decaulne A, Sæmundsson P. 2006. Geomorphic evidence for present-day snow-avalanche and debris-flow impact in the Icelandic Westfjords. *Geomorphology* **80**: 80–93.
- Decaulne A, Eggertsson Ó, Sæmundsson P. 2012. A first dendrogeomorphologic approach of snow avalanche magnitude-frequency in Northern Iceland. *Geomorphology* **167–168**: 35–44.
- Dubé S, Fillion L, Héty B. 2004. Tree-ring reconstruction of high-magnitude snow avalanches in the Northern Gaspé Peninsula, Québec, Canada. *Arctic, Antarctic and Alpine Research* **36**: 555–564.
- Gardner J. 1970. Geomorphic significance of avalanches in the Lake Louise area, Alberta, Canada. *Arctic and Alpine Research* **2**: 135–144.
- Gavrilă IG, Horvath C, Meseşan F, Pop O. 2018. *Snow avalanche susceptibility mapping using GIS in Parâng Mountains*, Proceedings of 34th Romanian National Symposium on Geomorphology and 19th Joint Geomorphological Meeting, Buzău.
- Germain D, Fillion L, Héty B. 2005. Snow avalanche activity after fire and logging disturbances, Northern Gaspé Peninsula, Quebec, Canada. *Canadian Journal of Earth Sciences* **42**: 2103–2116.
- Holmes RL. 1983. Computer-assisted quality control in tree-ring dating and measurement. *Tree-ring Bulletin* **43**: 69–78.
- Kogelnig-Mayer B, Stoffel M, Schneuwly-Bollschweiler M, Hübl J, Rudolf-Miklau F. 2011. Possibilities and limitations of dendrogeomorphic time series reconstructions on sites influenced by debris flows and frequent snow avalanche activity. *Arctic, Antarctic, and Alpine Research* **43**(4): 649–658.
- Meseşan F, Man TC, Pop OT, Gavrilă IG. 2019. Reconstructing snow-avalanche extent using remote sensing and dendrogeomorphology in Parâng Mountains, *Cold Regions Science and Technology* **157**: 97–109.
- Muntán E, García C, Oller P, Martí G, García A, Gutiérrez E. 2009. Reconstructing snow avalanches in the south-eastern Pyrenees. *Natural Hazards and Earth System Sciences* **9**: 1599–1612.
- Nyberg R. 1989. Observations of slushflows and their geomorphological effects in the Swedish Mountain area. *Geografiska Annaler Series A – Physical Geography* **71**: 185–198.
- Pop OT, Gavrilă IG, Roşian G, Meseşan F, Decaulne A, Holobacă IH, Anghel T. 2016. A century-long snow avalanche chronology reconstructed from tree rings in Parâng Mountains (Southern Carpathians, Romania). *Quaternary International* **415**: 430–440.
- Pop OT, Munteanu A, Meseşan F, Gavrilă IG, Timofte C, Holobacă, IH. 2017. Tree-ring-based reconstruction of high-magnitude snow avalanches in Piatra Craiului Mountains (Southern Carpathians, Romania). *Geografiska Annaler Series A – Physical Geography* **100** (2): 99–115.
- Pop OT, Holobacă IH, Răchită IG, Decaulne A, Hotea M, Horváth C. 2020. *Reconstitution dendrochronologique des avalanches de neige et conditions synoptiques associées à l'épisode avalancheux majeur de l'hiver 2005 dans les Monts Maramureş (Carpathes Orientales Roumaines)*. In: Bonnardot V, Quenol H, (eds.) Actes du XXXIIIe Colloque de L'Association Internationale de Climatologie, Rennes, France, 553–558.
- Rapp A. 1959. Avalanche boulder tongues in Lapland. *Geografiska Annaler* **41**: 34–48.
- Rayback SA. 1998. A dendrogeomorphological analysis of snow avalanches in the Colorado Front Range, USA. *Physical Geography* **19**: 502–515.
- Reardon BA, Pederson GT, Caruso CJ, Fagre DB. 2008. Spatial reconstructions and comparisons of historic snow avalanche frequency and extent using tree rings in Glacier National Park, Montana, USA. *Arctic, Antarctic and Alpine Research* **40**: 148–160.
- RINNTech. 2020. Technology for tree and wood analysis. <http://www.rinntech.de/index-28703.html>. Accessed Nov. 11, 2020
- Schaerer PA. 1972. Terrain and vegetation of snow avalanche sites at Rogers Pass, British Columbia B.C. *Geographical Series* **14**: 215–222.
- Schweingruber FH. 1996. *Tree Rings and Environment. Dendroecology*. Paul Haupt, Bern, Switzerland, 609 p.
- Shroder JF. 1978. Dendrogeomorphological analysis of mass movement, Table Cliff Plateau, Utah. *Quaternary Research* **9**: 168–185.
- Silhán K, Tichavský R. 2017. Snow-avalanche and debris-flow activity in the High Tatras Mountains: new data from using dendrogeomorphic survey. *Cold Regions Science and Technology* **134**: 45–53.
- Stoffel M and Bollschweiler M. 2008. Tree-ring analysis in natural hazards research – an overview. *Natural Hazards and Earth System Sciences* **8**: 187–202.

- Stoffel M, Bollschweiler M, Hassler GR. 2006. Differentiating past events on a cone influenced by debris-flow and snow avalanche activity – a dendrogeomorphological approach. *Earth Surface Processes and Landforms* **31**:1424–1437.
- Stoffel M, Butler DR, Corona C. 2013. Mass movements and tree rings: A guide to dendrogeomorphic field sampling. *Geomorphology* **200**: 106–120.
- Stoffel M, Corona C. 2014. Dendroecological Dating of Geomorphic Disturbance in Trees. *Tree-Ring Research* **70**(1): 3–20.
- Tichavský R, Silhán K, Stoffel M. 2017. Age-dependent sensitivity of trees disturbed by debris flows – Implications for dendrogeomorphic reconstructions. *Quaternary Geochronology* **42**: 63–75.
- Tumajer I, Trembl V. 2015. Reconstruction ability of dendrochronology in dating avalanche events in the Giant Mountains, Czech Republic. *Dendrochronologia* **34**: 1–9.
- Urdea P, Onaca A, Ardelean F, Ardelean M. 2011. *New evidence on the Quaternary glaciations in the Romanian Carpathians*. In: Ehlers J, Gibbard PL, Hughes PD (eds.), *Developments in Quaternary Science: Quaternary Glaciations – Extent and Chronology*. vol. 15. Elsevier, Amsterdam, The Netherlands, 305–322.
- Voiculescu M, Onaca A. 2013. Snow avalanche assessment in the Sinaia ski area (Bucegi Mountains, Southern Carpathians) using the dendrogeomorphology method. *Area* **45**:109–122.
- Ward RGW. 1985. Geomorphological evidence of avalanche activity in Scotland. *Geografiska Annaler Series A – Physical Geography*, **67**(3/4): 247–256.



---

## **In search of marine terraces from Central Dobrogea (Casimcea Plateau, Western Black Sea). Preliminary inquiries on distribution, extension and vertical crustal movements**

**Alexandru BERBECARIU<sup>1</sup>, Alfred VESPREMEANU–STROE<sup>1,2\*</sup>**

<sup>1</sup>Faculty of Geography, University of Bucharest, 1 Nicolae Bălcescu Blv., Bucharest 010041, Romania

<sup>2</sup>Sfântu Gheorghe Marine and Fluvial Research Station, Faculty of Geography, University of Bucharest, Sf. Gheorghe, Tulcea, Romania

Received 18 December 2020; Revised 24 December 2020; Accepted 24 December 2020

\*Correspondence to: Alfred VESPREMEANU–STROE; e-mail: [fredi@geo.unibuc.ro](mailto:fredi@geo.unibuc.ro)

---

### **ABSTRACT**

Casimcea Plateau is an uplifted (exhumated) peneplain cut in Proterozoic green–schists and one of the oldest tectonic units around the Black Sea. Despite its overall monotonous physiognomy, the plateau is crossed by Casimcea Valley and presents a seaward façade to the east which preserves (sub)horizontal surfaces as testimonies of the paleoenvironmental changes (sea level and climate). This research aims to identify the marine and fluvio–marine terraces and to define their vertical distribution based on the morphometric analysis of two study sites (north – Ceamurlia; south – Taşaul Lake) using EU-DEM. 6 levels were identified as possible marine terraces within the 2–50 m altitude range and also some inferences were made concerning the age of the lower three levels. Also, the present work highlights a differential (stronger) uplift of the northern sector between Peceneaga – Camena and Ostrov – Sinoe faults reflected by both the elevation difference of 5–6 m between the terraces staircases identified at the two sites and by the elevation gaps analysed on an array of cross-fault transects carried on over Ostrov – Sinoe fault.

**KEYWORDS** Casimcea Plateau, morphometry, marine terraces, faults, uplift

---

### **1. Introduction**

A marine terrace is any relatively flat, horizontal or gently inclined surface of marine origin, bounded by a steeper ascending slope on one side and by a

steeper descending slope on the opposite side. In temperate regions marine terraces often result from marine erosion (abrasion or denudation) (marine-cut terraces or shore platforms) or consist of shallow-water to slightly emerged accumulations of materials

removed by shore erosion (marine-built terraces) (Pirazzoli, 2005). Marine terraces are unique features which testify the long term geomorphological evolution in relation with paleoclimate and sea level oscillations.

The presence and spatial distribution of marine terraces across the Romanian coast of the Black Sea is a poorly-known issue despite the old debate concerning their existence and the frequency/number (Posea, 1980, 1983; Ielenicz, 1988). Most of these studies tried to identify the supposed marine terraces and to describe their physiognomy and spatial distribution but, at the same time uncertainty about their existence persists as long as no study proved this with clear evidence.

De Martonne (1921) identified three possible abrasion terraces in northern Dobrogea, whereas Brătescu (1943) concluded that marine terraces have been entirely submerged during the Holocene transgression. Latter, Roșu (1969) argued on the existence of marine terraces extending inland along the Danube and Danube Delta, identifying up to 7 abrasion levels (2–4 m; 8–10 m; 15–20 m; 30–45 m; 55–65 m; 75–85 m; 95–110 m). Ielenicz (1988) developed the previously proposed hypothesis (Posea, 1983) and concluded on the existence of four marine terraces levels (3–5 m, 6–8 m, 10–15 m, 20–30 m) along the present-day coastline.

The present study deals with the morphometric analysis of the lower reach of Casimcea Valley (Tașaul Lake basin) and the north-east façade of the Casimcea Plateau which fronts presently the Ceamurlia–Golovița–Zmeica–Sinoe lagoons.

The uplifting pattern associated with marine terraces is a subject of discussion in tectonically active basins. Within the Mediterranean basin, the association of marine terraces with the collision between Eurasian Plate and African Plate largely acknowledged, this factor being related to both seismic movements and coastal uplifting patterns (Cucci and Cinti, 1998; Zazo et al., 1999, 2003; Zecchin et al., 2004; Pirazzoli et al., 2004). On the other hand, although the relationship between the marine terraces extent, spatial distribution and vertical crustal movements along the Romanian Black Sea coastline is not well documented, this subject was extensively discussed within the southern Black Sea basin by Keskin

et al. (2011), Yildirim et al. (2013), Mc Clain et al. (2016), Berndt et al. (2017), Spencer et al. (2017), Erturac et al. (2019; 2020) by analysing the terraces from northern Turkey, associated with the neo-tectonic and seismic movements related to the North Anatolian Fault.

The aim of this study is to identify and characterize those morphological features which could account for the existence of marine terraces in central Dobrogea. Additionally, we inspected the elevation differences between the identified terrace levels trying to assess the contribution of the vertical crustal movements, associated with some of the major tectonic faults based on the elevation differences between terrace levels.

## 2. Regional settings

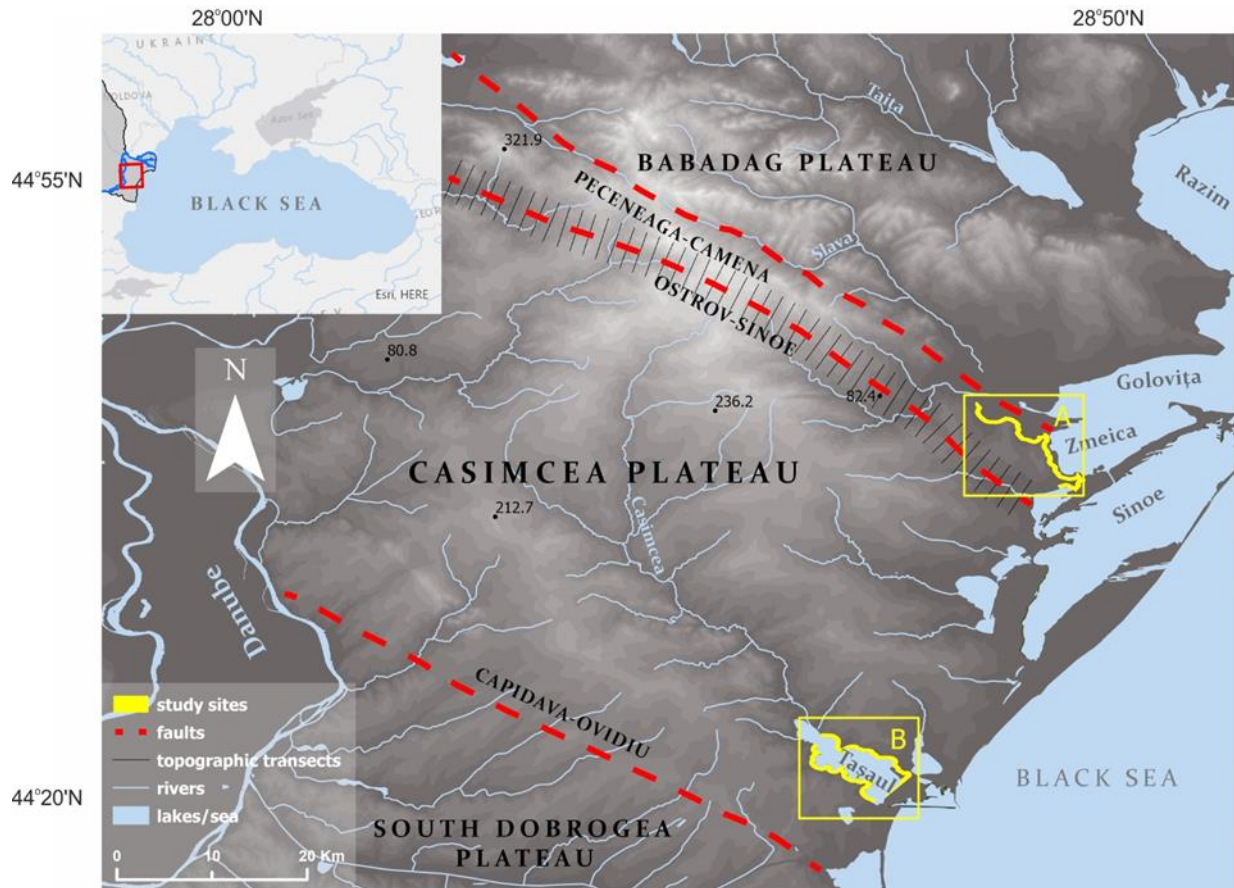
Our study area overlays the Casimcea Plateau (Fig. 1), one of the most affected relief units around the Black Sea by the processes of marine transgression and regression during the geological-time due to its long lasting evolution (since Proterozoic; Săndulescu, 1984) and slightly uplifting. (Rădulescu et al., 1976 and Visarion et al., 1988). The Casimcea Plateau is placed between Babadag Plateau (N) and South Dobrogea Plateau. Peceneaga – Camena and Capidava – Ovidiu faults disrupt the plateau to the N and to the S rendering it a perched exhumated peneplain as it was affected by tectonic uplifting (Ciocârdel and Esca, 1966; Polonic et al., 1999).

Late Pleistocene loess deposits prevail, including silt or clayey silt and isolated Pleistocene calcareous pediments along the western banks of the Ceamurlia–Golovița–Zmeica–Sinoe Lagoon Complex. The geological features around Tașaul Lake are more heterogeneous: the Precambrian green-schists are predominant on the western bank and the Late Pleistocene loessoid deposits are predominant on the other side of the lake, interfering with Pleistocene calcareous layers.

As one of the oldest geological and geomorphological unit in western Black Sea, Casimcea Plateau displays a complex morphology with very steep vertical walls, terraces, canyons superimposed on a widely developed peneplaine with mild slopes resulted from long time subaerial evolution (Posea, 1980; Comănescu, 2004). This complex morphology

is also imposed by the heterolytic composition with outcropping green schist, limestones or loess (Comănescu, 2004, Ielenicz et al., 2004) and their exposure to multiple shaping factors (i.e. tectonic, subaerial, coastal and marine) which acted at multiple time scales (Ielenicz, 1988). The genesis of marine terraces is generally associated with the highstands

of sea level, which occurred during the odd Marine Isotope Stages (MIS) specific to warm (inter-glacial) periods from MIS 1, MIS 3, MIS 5a and MIS 5e to MIS 15 (Yildirim, 2013); during all these intervals the Black Sea basin was connected with the Mediterranean Sea, except MIS 3 (Badertscher et al., 2011).



**Figure 1** Location map showing the position of the study sites in the western Black Sea and in the Casimcea Plateau: Ceamurlia–Golovița–Zmeica–Sinoe Lagoon Complex (A) and Tașaul Lake (B) in the context of the major units of relief from Central Dobrogea

The Casimcea Plateau is mainly drained by Casimcea River, which is the most important river in region, excluding Danube, which crosses Central Dobrogea following an old syncline alignment from NW to SE parallel to the main faults. It sums up a drainage basin of 740 km<sup>2</sup>, a length of 69 km and a multiannual average liquid flow of 0.65 m<sup>3</sup>/s at Casian hydrometric station (Zaharia and Pișota, 2003). The present-day climate is dry–temperate, with mean annual temperature of ca. 11 °C and mean precipitation of ca. 400 mm/yr (Sandu et al., 2008). Most of this occurs

during summer as heavy rainfalls within relatively short time intervals. This has a relatively high impact in sediment mobilization and subsequent morphological changes.

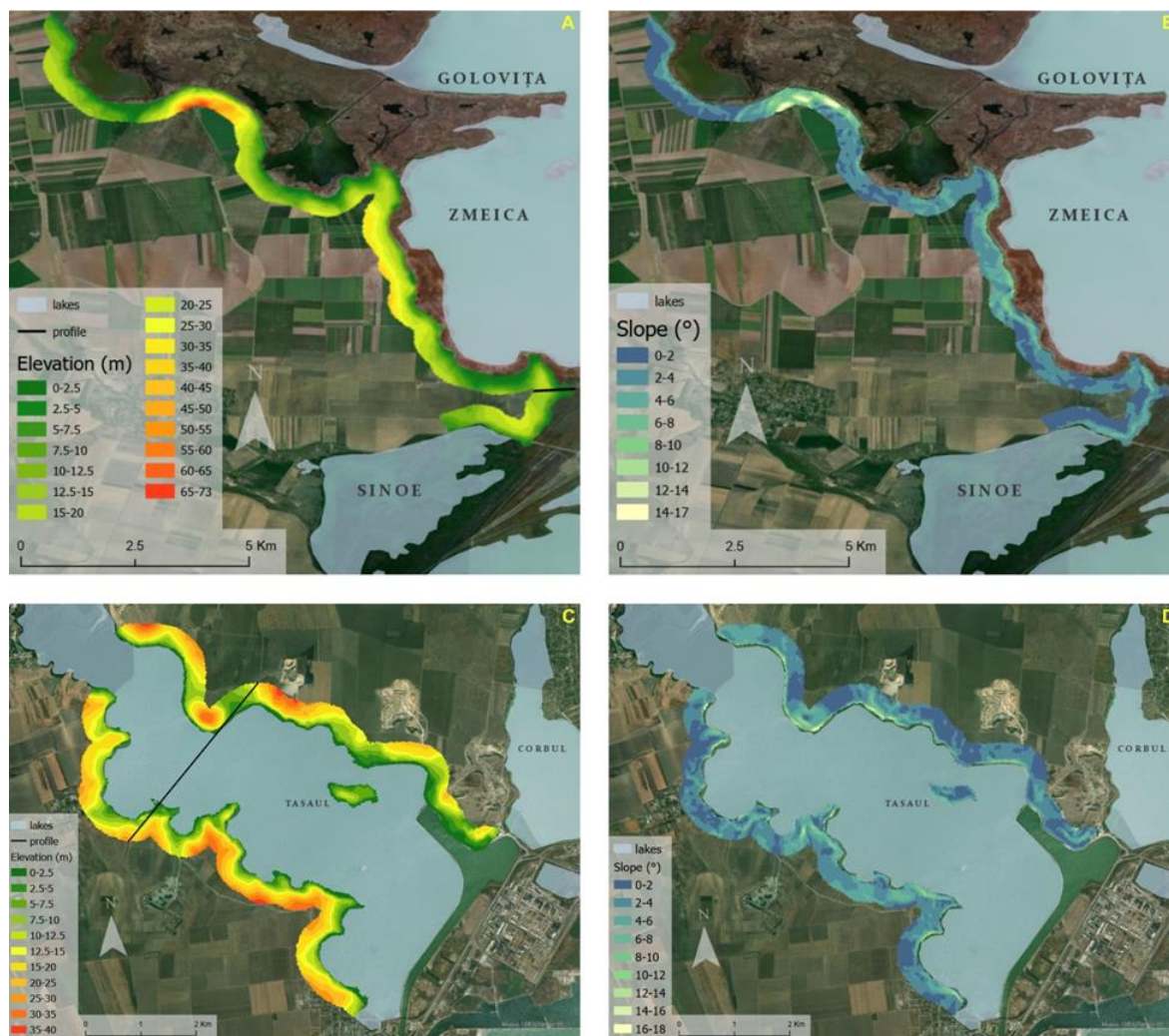
### 3. Methodology

For the hypsometry and slope analysis we used the Digital Elevation Model over Europe (EU-DEM) with a spatial resolution of 25 m and a vertical resolution of ±7 m (www.eea.europa.eu). Two polygons were created for the two study sites, both with a constant

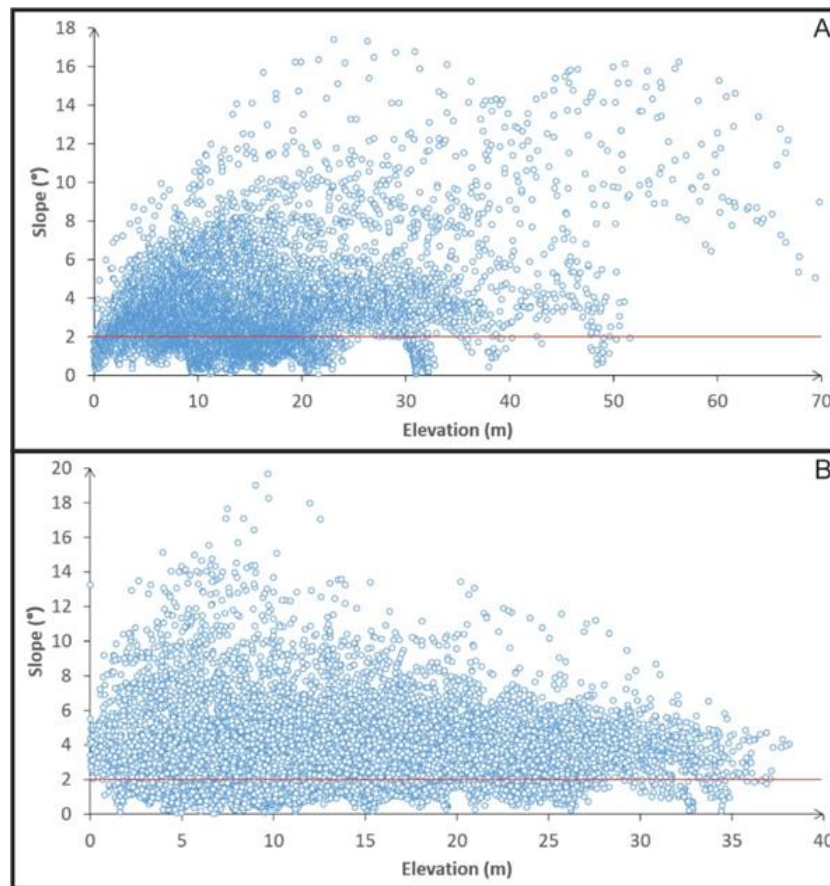
width of 400 m drawn upslope from the waterlines of the Taşaul Lake (but excluding the Casimcea river delta and the littoral barrier) and from the western waterlines of the Ceamurlia–Golovița–Zmeica–Sinoe lagoons. The slope values were distributed per altitude classes of 2 m elevation and reclassified to select and analyse the sub-horizontal ( $1-2^\circ$ ) and horizontal surfaces ( $\leq 1^\circ$ ); these values were associated with the hypsometric distribution by transforming the EU-DEM raster to points. All the resulted slope and elevation associated points were joined by using spatial analysis intersection tool in ArcGis Pro software; Arc Map 10.8, Global Mapper 15 and Google Earth Pro were used for identifying the areas with quasi-horizontal levels and for creating ranges of elevation profiles. The geological settings of our study sites and

the position of the main faults (Pecenaga–Camena; Ostrov–Sinoe; Capidava–Ovidiu) were inspected on 1:50000 geological maps for this areas (153 d Jurilovca; L-35-118-D and 169 c Gura Dobrogei L-35-130-C).

The analysis concerning the vertical crustal movements of the areas situated north and south of the Ostrov – Sinoe fault comprised 53 transects (with the equidistance of 1-km) placed normal on the Ostrov – Sinoe fault. Each transect had a 4-km length with the mid-point right on the fault. Further, the cells (from the transects) placed at the same distance from the fault (e.g. 0–25 m, 25–50 m etc.) were averaged to get a cross-fault hypsometry pattern synthesized in the cross-fault averaged profile.



**Figure 2** Elevation and slope maps of the two study sites from eastern Casimcea Plateau. A. Hypsometry Ceamurlia site); B. Slopes (Ceamurlia site); C. Hypsometry (Taşaul Lake); D. Slopes (Taşaul Lake)



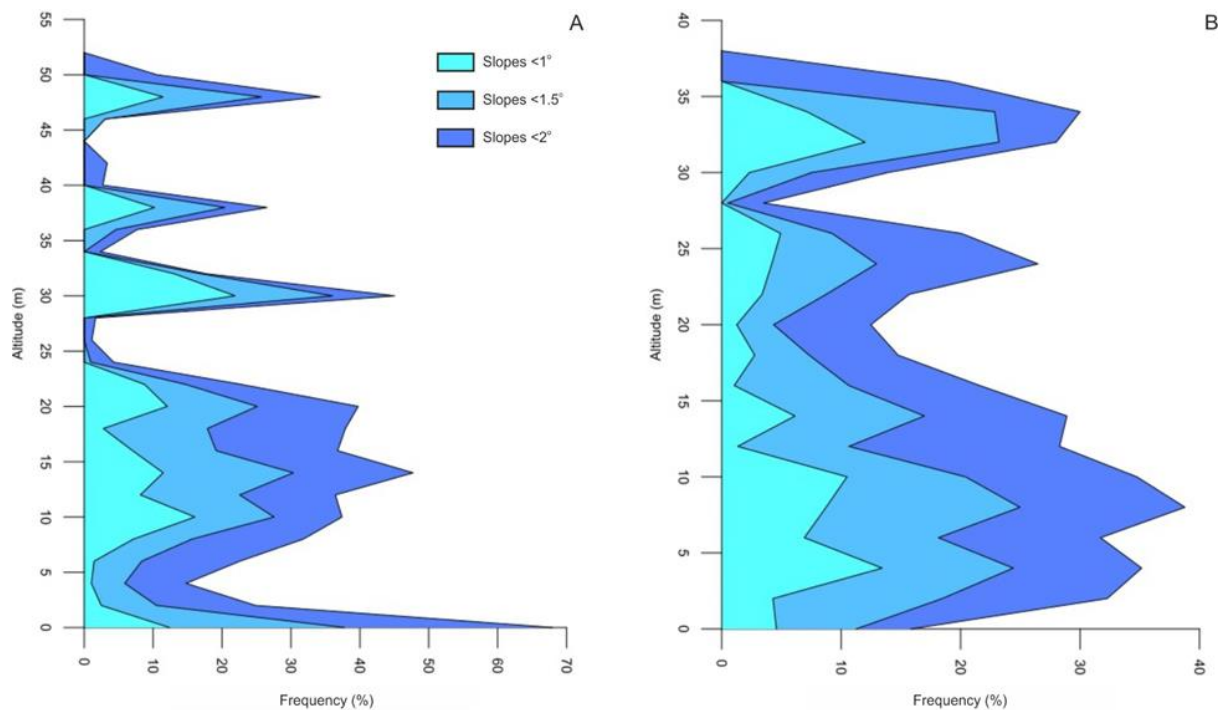
**Figure 3** Co-distribution (scattered) of slopes and elevation in Ceamurlia (A) and Taşaul Lake (B)

#### 4. Results

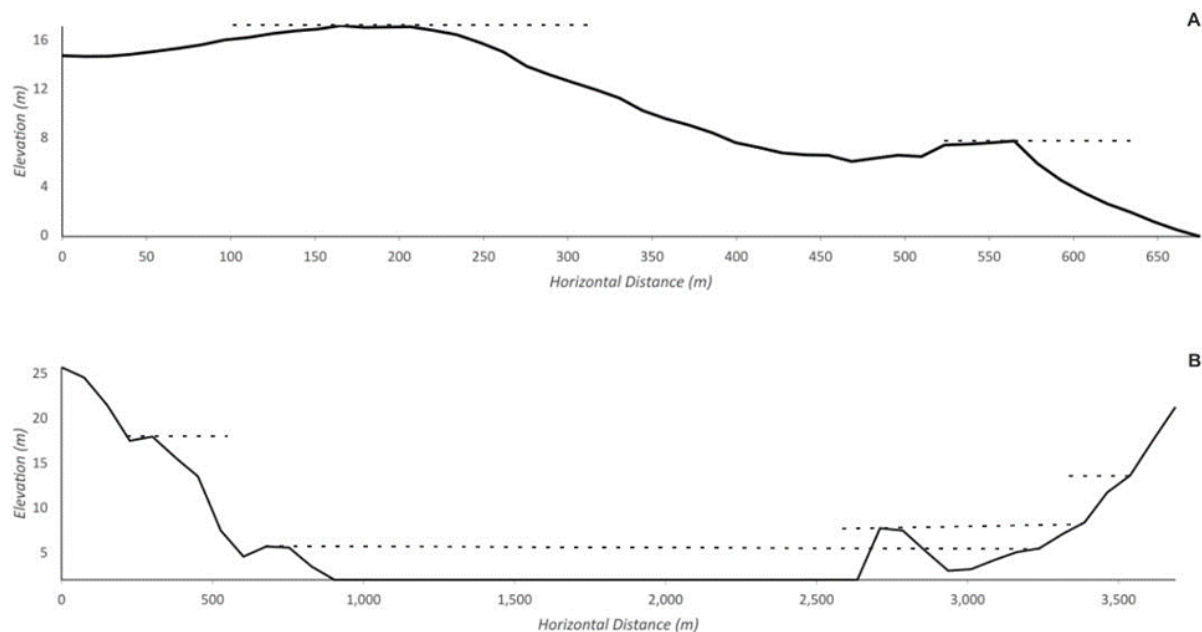
The spatial repartition of both elevation and slope is a suitable method that helped us to find the sub-horizontal levels which are susceptible to be marine terraces (Fig. 2). At our study sites which extend over a surface area of 312 km<sup>2</sup> (Taşaul) and 288 km<sup>2</sup> (Ceamurlia), the slope ranges between 0° and 19.6° for Taşaul and between 0° and 17.4° for Ceamurlia. In both areas, the steeper surfaces sit in the proximity of waterlines, followed by a transition to gentler slopes further inland. The sub-horizontal surfaces with slopes <2° and <1.5° represent 27.09%, respectively 15.22%, whilst the horizontal areas (with a slope <1°) total 6.15% for Taşaul. In the northern-exposed study site of Ceamurlia the frequency of the sub-horizontal surfaces is of 31.82% (with a slope <2°), respectively 18.14% (with a slope <1.5°). Regarding the horizontal areas with slopes <1°, they occupy 7.40%.

The elevation distribution in the two sites is roughly similar but with differences concerning the

maximum altitudes. Thus, the area surrounding Taşaul Lake displays a range of heights between 0 m and 39.8 m, while the Ceamurlia hypsometry unfolds between 0 m to 73.1 m. Nevertheless, even if the elevation ranges are different in these two areas, it can be observed the same pattern of slopes distribution as shown by the joint analysis of elevation and slope (Fig. 3). The distribution of the sub-horizontal (1–2°) and horizontal (<1°) surfaces shows a very distinct pattern with a significant clustering within the following altitude intervals: 2–5 m, 7–11 m, 13–16 m, 22–27 m, 31–35 m for Taşaul and 0–2 m, 7–11 m, 13–16 m, 19–23 m, 28–33 m, 36–39 m and 46–49 m for Ceamurlia (Fig. 4). An array of cross-slope profiles which generally shows one to three steps on the same linear profile reveals the presence and the aspect of the quasi-horizontal levels. Figure 5 shows two topographic profiles that intercepted the levels of 2–5, 7–11, 13–16 m for Taşaul and the levels of 7–11 and 13–16 m for Ceamurlia.



**Figure 4** The distribution of elevation–slope within Ceamurlia (A) and Taşaul Lake (B) study sites



**Figure 5** Selected topographic profiles showing the distribution of (quasi)horizontal surfaces from the two study sites. Two levels are indicated within the Ceamurlia site (A), while four were identified in the proximity of the Taşaul Lake (B)

## 5. Discussion

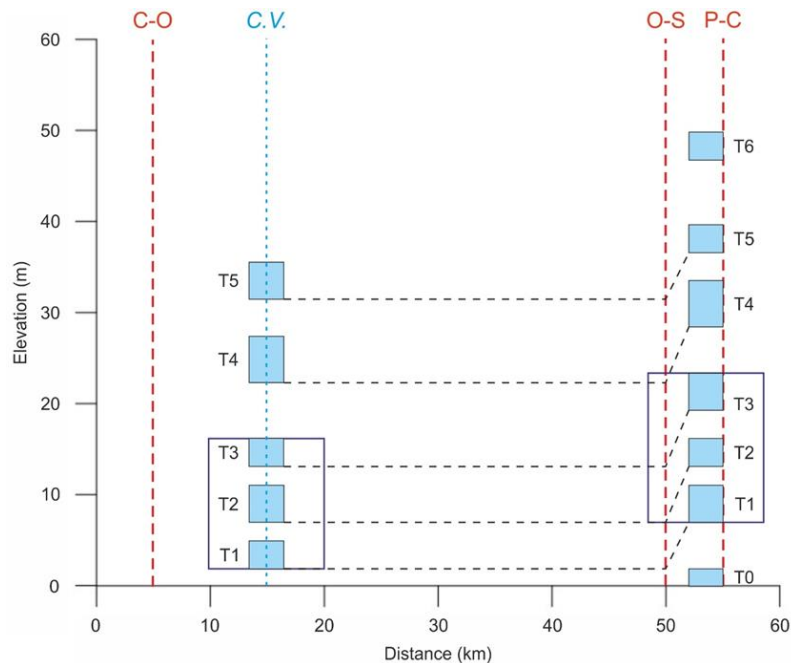
The two study sites have a common convex cross-slope profile which is specific to areas affected by lateral/horizontal erosion stronger than the overall vertical erosion; the latter one is derived from weathering, sheet-wash and rill-wash (including gullies and

torrents). In these cases, the extra-erosion which acted horizontally and imposed both the upward slope convex profile but also the incision of (sub)horizontal steps at different altitudes seems to be mainly of marine origin (abrasion and hydraulic pressure) especially for the north-eastern façade of Casimcea

Plateau (Ceamurlia) and of combined origin – marine and fluvial – for the present Taşaul Lake slopes. Due to its frontal exposure to the north–eastern storm waves, Ceamurlia probably developed most of the (sub)horizontal steps during the intervals of Black Sea highstands when the marine waters could have occupied part of the territory which accommodates now the delta lagoons. On the present lower reach of Casimcea Valley, the reconnection to the Mediterranean Sea (ca 9400 BP, Ankindinova et al., 2020) and the subsequent early and mid-Holocene transgression drowned the floodplain and the base of the lateral slopes and transformed it firstly into a ria and later into a deep liman as suggested by the deep water of the present Taşaul Lake and by the thick Holocene sediments strata (>30 m, as indicated by drills) covering the former Casimcea floodplain within the present coastal barrier. The mouth of Casimcea River was closed by a sandy barrier by the time of the sea–level rise slowdown (ca. 5000 BP) as most of the coastal barriers from the western Black Sea (Caraivan et al., 2015; Vespremeanu–Stroe et al., 2016, 2017). The (sub)horizontal steps cut in the lateral slopes of the Taşaul Lake most probably are fluvial terraces which evolution was imposed by the high position recorded by the Black Sea which acted as a base level for the

Casimcea River. It is reasonable to assume that some of these steps were also shaped by direct marine (ria) or liman water dynamics which shaped them as fluvio–marine terraces.

Our analysis highlights the presence of at least 6 (sub)horizontal steps cut into the slopes of the two study sites which probably are (all or most of them) marine or fluvio–marine terraces. Even though, the sites are located in different (north and south) sectors of the Casimcea Plateau at a relatively long distance (40 km) apart, the spatial (vertical) pattern of their distribution is near identical with the same vertical development of each terrace and the same vertical space in-between the terraces. This coherent pattern made us assuming that they have the same modelling agent of marine origin (e.g. waves). The only significant difference is the fact that terraces from the northern site are ca. 5–6 m higher than their correspondents from the southern site (Taşaul Lake) (Fig. 6). Noteworthy, the constant vertical difference for all terraces could be explained by the Ostrov – Sinoe fault line (re)activation after the development of the lower terrace and the vertical displacement of all the steps from Ceamurlia site (T1–T6) excepting the lowermost one (T0).

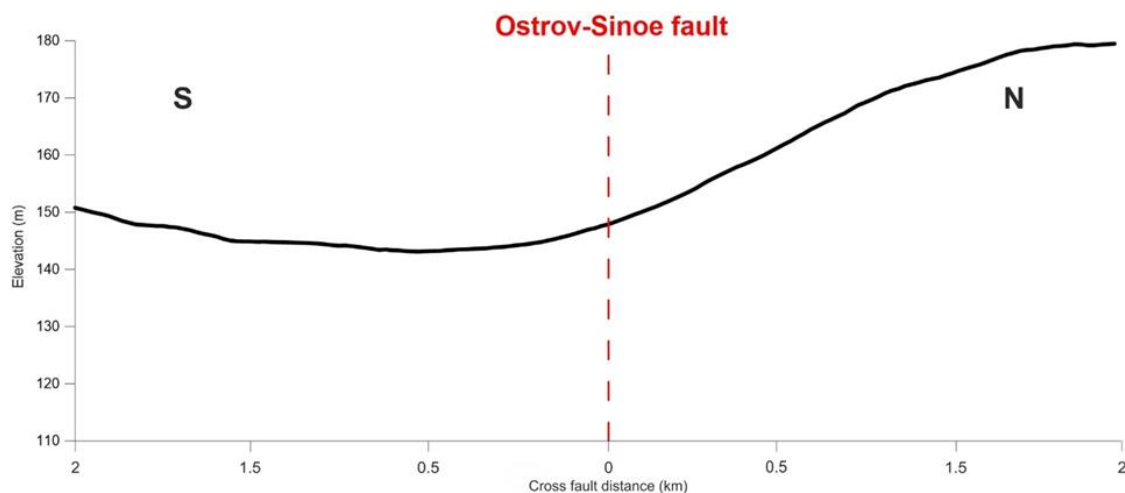


**Figure 6** The altitudinal correlation between the possible marine terraces from the two study sites: C-O = Capidava-Ovidiu crustal fault; C.V. = Casimcea Valley; O-S = Ostrov-Sinoe fault; P-C = Peceneaga-Camena crustal fault

In the absence of field evidence (stratigraphy and chronology) is complicated to infer the age of these marine terraces. Still, most probably their formation is linked to the highest and the most long-lasting highstands of Black Sea from Middle and Late Pleistocene if we take into account the age (MIS15; ca. 600 ka) established for the oldest marine terraces from Black Sea (Keskin et al., 2011; Yildirim et al., 2013).

Faunal evidence for water intrusions from the Mediterranean Sea into the Black Sea during the last 600 ka (MIS 15 – present) shows a minimum of 6 intervals (Svitoch et al., 2000) and a maximum of 9 intervals (Zubakov, 1988) when the two seas were connected. But a more recent study reconstructing the reconnection history based on oxygen isotope ( $\delta^{18}O$ ) signatures from a coastal cave (Sofular) provides evidence for at least twelve time intervals within the last 670 ka when water exchange between the Black Sea and Mediterranean Sea was established which should coincide with sea levels higher than the current Bosphorus sill depth of  $-35$  m b.s.l. (Bardetscher et al., 2011). In our case, the 0–2 m terrace (T0) from the northern site should be middle and late

Holocene as it corresponds with the current sea-level stabilization (5000 BP – present). The next three terraces levels, T1–T3, developed at small vertical distances (ca. 3 m) in-between them and the lowermost one (T1) seems to be formed either during MIS 3 (41–30 ka) when terrace formation was also reported for the Sakarya river (Erturac et al., 2019) or MIS 5a (84–72 ka) when the lowermost terrace was incised in Sinop region and Adapazari basin, Anatolia (Yildirim et al., 2013; Erturac et al., 2020). Next levels of terraces are even more difficult to assess in the absence of absolute dating, but with a possible correspondence of T2 or T3 (which are the best developed in Casimcea Plateau; highest frequencies) to MIS 5e (125–119 ka) when the last higher than present World Ocean level (0–10 m a.s.l.) was reported (Pirazzoli, 1991; 1996). The vertical crustal movements from Central Dobrogea were documented and mapped by several authors between 1966 and 1999 (Ciocârdel and Esca, 1966; Cornea et al., 1979; Polonic et al., 1999, cited in Dimitriu and Sava, 2007), which largely considered Dobrogea either quasi-stable or slightly uplifting.



**Figure 7** Cross-fault averaged topographic profile (4 km long) intersecting Ostrov-Sinoe fault line showing the uplifting (N) and the lateral compartments

More recently, the emplacement of several GNSS stations (in 2013) enables permanent vertical movement monitoring which for this short time interval presents subsidence at rates between  $-10$  mm/yr (Babadag) and  $-6$  mm/yr (C. Midia) across the coastal strip of our study site (Dimitriu et al., 2017). Still, these values must be considered with caution due to the

very short time of analysis (4 years) and the errors which could be associated with the station emplacement especially during first years when the terrain may compact and settle. Nevertheless, an uplifting tendency between Peceneaga – Camena and Capi-dava – Ovidiu crustal faults is firstly suggested by Rădulescu et al. (1976) and Visarion et al. (1988), cited

in Biter et al. (1998), but also by Diaconescu et al. (2019) especially for the northern unit of Casimcea Plateau between the Peceneaga – Camena crustal fault and the Ostrov – Sinoe strike-slip fault. The south–north cross-fault averaged profile carried on in the present study for the Ostrov – Sinoe fault shows vertical differences up to 30 m depending on the cross-fault distances (it varies from 150.8 m in south to 179.5 m in the north; Fig. 7). The evolution patterns of faults and, generally, the crustal movements are subjects that need more complex analysis. Still, our results prove an uplifted activity north of Ostrov – Sinoe fault, for the compartment placed between the Ostrov – Sinoe and Peceneaga – Camena faults, which could be also responsible for the vertical gap between the altitudinal distribution of terraces from the two Taşaul Lake and Ceamurlia sites (Fig. 6).

## 6. Conclusions and future work

The present geomorphological analysis identified 7 (sub)horizontal surfaces, with the lowermost (0–2 m) corresponding to the late Holocene Black Sea level which suggests the probable presence of 6 marine terraces within the 2–50 m elevation range. The lower terraces (T1–T3) were most probably shaped during MIS 5 (MIS 5a, MIS 5e) but MIS 3 cannot be excluded as an alternative age for the lowermost one (T1). Our study also proves a differential (stronger) uplift of the northern sector between Peceneaga – Camena and Ostrov – Sinoe faults which is indicated both by the elevation difference of 5–6 m between the terraces staircases identified in the two sites and by the elevation gaps analysed on an array of cross-fault transects carried on over Ostrov – Sinoe fault.

Further work is needed to confirm and develop these preliminary results: it is necessary both an in-depth morphometric analysis carried on a higher accuracy DEM (with centimetre resolution) using drones and/or terrestrial laser scan and sedimentological and chrono-stratigraphical analyses to describe and date the deposits which overlay the marine terraces.

## Acknowledgements

The authors are grateful to Dr. Luminița Preoteasa for her devoted help regarding the process of reviewing

and improving this manuscript with relevant scientific information and also to Laurențiu Țuțuianu for his valuable help concerning the graphs conception. We would like to thank Dr. Nicolae Cruceru for providing us useful suggestions in terms of Romanian geographical literature. This contribution is part of Sfântu Gheorghe Marine and Fluvial Research Station (Faculty of Geography, University of Bucharest) research programme.

## References

- Ankidinova O, Aksu A.E, Hiscott R.N. 2020. Holocene sedimentation in the southwestern Black Sea: Interplay between riverine supply, coastal eddies of the Rim Current, surface and internal waves, and saline underflow through the Strait of Bosphorus. *Marine Geology*, Volume **420**, February 2020. 106092.
- Badertscher S, Fleitmann D, Cheng H, Edwards R.L, Göktürk O.M, Zumbühl A, Leuenberger M, Tüysüz O. 2011. Pleistocene water intrusions from the Mediterranean and Caspian seas into the Black Sea. *Nature Geoscience*, **4**: 236–239.
- Berndt C, Yildirim C, Çiner A, Ertunç G, Sarikaya M.A, Özcan O, Güneç Kiyak N, Öztürk T. 2017. Timing and development of Late Quaternary fluvial terraces of the lower course of Kızılırmak River (Northern Turkey). Conference Proceedings: EGU 2017. Vienna. Austria.
- Biter M, Malita Z, Diaconescu M, Rădulescu F, Nacu V. 1997. Crustal movements and earthquakes distribution in Dobrudja and Black Sea. National Institute for Earth Physics, Bucharest Măgurele. P.O. Box MG-2, Romania.
- Brătescu C. 1943. Oscilațiile bazinului Mării Negre în Cuaternar. *BSRRG*. vol **I** (LXI).
- Caraivan G, Opreanu P, Voinea V, Iulian P, Giosan L. 2015. Holocene landscape changes and migration of human communities in the western part of the Black Sea (Mamaia Lake area). **IGCP 610** Third Plenary Conference.
- Ciocârdel R, Esca A. 1966. Essai de synthese des donnees actuelles concernant les mouvements verticaux recents de l'ecorce terrestre en Roumanie: *Rev. Roum. de Geol. Geophys. et Geogr., Serie de Geophysique*, v. **10**, n. 1.
- Comănescu L. 2004. *Bazinul morfohidrografic Casimcea: studiu geomorfologic*. Editura Universitatii din Bucuresti, București.
- Cornea I, Drăgoescu I, Popescu M, Visarion, M. 1979. Harta mișcărilor crustale verticale recente de pe teritoriul R.S. România. *St. Cerc. Geol. Geof. Geogr., Seria Geofizică*, v. **17**, p. 3–20.

- Cucci L, Cinti F.R. 1998. Regional uplift and local tectonic deformation recorded by the Quaternary marine terraces on the Ionian coast of northern Calabria (southern Italy). *Tectonophysics*, **292**(1–2).
- Diaconescu M, Craiu A, Toma–Danila D, Craiu G.M. 2019. Main active faults from the eastern part of Romania (Dobrogea and Black Sea). part I: Longitudinal faults system. *Romanian Reports in Physics*. **71**(1): 702.
- Dimitriu R.G, Sava C.S. 2007. Considerații asupra proceselor geodinamice actuale din Dobrogea – o pledoarie în favoarea realizării „poligonului geodinamic Dobrogea”. INCD GeoEcoMar. București.
- Dimitriu R.G, Stanciu I.M., Barbu M–B, Dobrev N, Dumitriu P.D. 2017. The first results on the western Black Sea coastal geodynamics resulted from geopotential permanent GNSS stations network data processing. *17th International Multidisciplinary Scientific Geoconference*, Albenă. Bulgaria. 149–157.
- Erturac K.M, Şahiner E, Sağlam S.A, Gürbüz A, Okur H, Zabcı C, Meriç N, Özeren S, Gürsel S. 2020. Spatio-temporal variations on the vertical deformation rate of the NW Anatolian Block: Luminescence chronology of the Sakarya River terraces. 22nd EGU General Assembly, held online 4–8 May, 2020.
- Erturac K.M, Şahiner E, Forman S.L, Kazancı N. 2019. Combination of fluvial and coastal records of the southern Black Sea basin to reveal sea-level changes during the Late Pleistocene. Conference: INQUA. Dublin. Ireland.
- Ielenicz M, Comănescu L, Nedelea A. 2004. Loess-related landforms in central and northern Dobrogea – Romania. *Revista de Geomorfologie*, nr. **6**. Editura Universității București, p. 27–34.
- Ielenicz M. 1988. Terasale din Dobrogea. *Analele Universității din București*, **XXXIX**.
- Keskin S, Pedoja K, Bektaş O. 2011. Coastal uplift along the eastern Black Sea coast: New marine terrace data from Eastern Pontides, Trabzon (Turkey) and a review. *Journal of Coastal Research*, **27**(6a): 63–73.
- Martonne E. de. 1924. Excursions géographiques de l'Institut de géographie de l'Université de Cluj. În *Lucrările Institutului Universității din Cluj*, vol **I**.
- Mc Clain K, Şahina S, Yildirim C. Çinera A, Sarıkaya M.A, Öztürk T, Güneç Kiyak N. 2016. Quantification of fluvial response to tectonic deformation and climate change in the Central Pontides; Inferences from OSL dating of fluvial terraces. Conference: 69th Geological Congress of Turkey. Ankara. Turkey.
- Pirazzoli P.A, Laborel J, Stiros S.C. 1996. Earthquake clustering in the eastern Mediterranean during historical times. *Journal of Geophysical Research: Solid Earth*, **101**(B3), 6083–6097.
- Pirazzoli P.A, Stiros SC, Fontugne M, Arnold M. 2004. Holocene and Quaternary uplift in the central part of the southern coast of the Corinth Gulf (Greece). *Marine Geology*, **212**: 35–44.
- Pirazzoli P.A. 1991. *World atlas of Holocene sea-level changes*. Elsevier oceanography series. Vol **58**, pp 1–280; ref: 41 p.1/2.
- Polonic G, Zugrăvescu D, Horomnea M, Dragomir V. 1999. Crustal vertical recent movements and the geodynamic compartments of Romanian territory: Second Geophysical Congress, Istanbul. Book of Abstracts: p. 300–301.
- Posea G. 1980. Teras marine în Dobrogea? *Terra*, XII (XXXII) 3.
- Posea G. 1983. *Terasale lacustre și marine*. vol. I. Geografie Fizică. Editura Academiei Române. București.
- Rădulescu D.P, Cornea I, Săndulescu M, Constantinescu P, Rădulescu F, Pompilian A. 1976. Structure de la croute terrestre en Roumanie. Essai d'interprétation des études sismiques profondes. *An. Inst. Geol. Geofiz.*, **50**: 5–36.
- Roşu A. 1969. Observații geomorfologice pe latura de nord a Dobrogei. În vol. „*Studii geografice asupra Dobrogei*”. **XIII–XIV**.
- Sandu I, Pescaru V.I, Poiana I. 2008. *Clima României*. Editura Academiei Române. București.
- Săndulescu M. 1984. *Geotectonica României*. Ed. Tehnică. București. 333p. (in Romanian)
- Schwartz M. 2005. *Encyclopedia of Coastal Science*. Pirazzoli P.A. Marine Terraces. p. 632. Springer.
- Spencer J, Emre T, Sözbilir H, Turan M. 2017. Late Quaternary rapid uplift deduced from marine terraces in Eastern Pontides, Turkey. The Geological Society of America, GSA annual meeting. Seattle. Washington, USA.
- Svitoch A.A, Selivanov A.O, Yanina T.A. 2000. Paleohydrology of the Black Sea Pleistocene basin. *Wat. Resour.*, **27**: 655–664.
- Vespremeanu–Stroe A, Preoteasa L, Zăinescu F, Rotaru S, Croitoru L, Timar–Gabor A. 2016. Formation of Danube delta beach ridge plains and signatures in morphology. *Quaternary International*, **415**: 268–285.
- Vespremeanu–Stroe A, Zăinescu F, Preoteasa L, Tătui F, Rotaru S, Morhange C, Stoica M, Hanganu J, Gabor–Timar A, Cărdan I. 2017. Holocene evolution of the Danube delta: An integral reconstruction and a revised chronology. *Marine Geology*, **388**: 38–61.
- Visarion M, Săndulescu M, Stănică D, Veliciu S. 1988. Contributions a la connaissance de la structure profonde de la Plate-forme Moesienne en Roumanie. *St. tehn. și econ. Seria D*, **15**, Geofizică. București.
- Yildirim C, Melnick D, Ballato P, Schildgen T.F, Echtler H, Evren Erginal A, Güneç Kiyak N, Strecker M.R. 2013. Differential uplift along the northern margin of the Central Anatolian Plateau: Inferences from marine terraces. *Quaternary Science Reviews*, Volume **81**, pages 12–28.

- Zaharia L, Pişota I. 2003. Dobrogea – 1. Apele Dobrogei. p. 107. *Analele Universităţii din Bucureşti, Geografie*.
- Zazo C, Goy J. L, Dabrio C.J, Bardaji T, Hillaire–Marcel C, Ghaleb B, Soler V. 2003. Pleistocene raised marine terraces of the Spanish Mediterranean and Atlantic coasts: records of coastal uplift, sea–level highstands and climate changes. *Marine Geology*, **194**(1–2): 103–133.
- Zazo C, Silva P.G, Goy J.L, Hillaire–Marcel C, Ghaleb B, Lario J, Bardajii T, Gonzalez A. 1999. Coastal uplift in continental collision plate boundaries: data from the Last Interglacial marine terraces of the Gibraltar Strait area (south Spain). *Tectonophysics*, **30**: 95–109.
- Zecchin M, Nalin R, Roda C. 2004. Raised Pleistocene marine terraces of the Crotona peninsula (Calabria, southern Italy): facies analysis and organization of their deposits. *Sedimentary Geology*, **172**(1–2).
- Zubakov V.A. 1988. Climatostratigraphic scheme of the Black Sea Pleistocene and its correlation with the oxygen isotope scale and glacial events. *Quat. Res.*, **29**: 1–24.
- \*\*\*<https://www.eea.europa.eu/data-and-maps/data/eu-dem>
- \*\*\*Harta Geologică a României. Scara 1:50000. Institutul de Geologie şi Geofizică.