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LATE PLEISTOCENE GLACIATION IN THE HEADWATERS OF THE CEREMUȘUL ALB VALLEY (MARAMUREȘ MOUNTAINS, ROMANIA)

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Abstract

The Late Pleistocene Jupania palaeoglacier (area 0.85 km^2, 1.7 km long) was reconstructed in the headwaters of the Ceremușul Alb/Bilyj Cheremosh valley (Maramureș Mountains). The study area represents one of the most inaccessible natural areas in the Romanian part of the Eastern Carpathians where the legacy of the Pleistocene glaciation has recently been discovered. Based on mapping of glacial landforms and deposits, we reconstruct glacier dimension and ice-surface geometry, as well as estimate equilibrium line altitude (ELA) during the maximal ice extent (MIE). Well-preserved terminal moraines mark the extent of glacier front at ~1400 m a.s.l. Sedimentological analysis documents that the lateral moraines are sometimes overbuilt by 1-1.5 m thick colluvial deposits. The ELA for the Jupania palaeoglacier calculated with the Area-Altitude-Balance-Ratio (AABR) 1.6 was 1630 m. However, the gentle-sloping mountain-top could serve as an important snow contribution area to glacier mass balance; therefore, the ELA could potentially exist even higher at 1676 m. The resulting climatic ELA (1630-1676 m) in the south-eastern part of the Maramureș Mountains fits well with the rising trend of ELA towards the southeast observed between Chornohora (ELA = 1516 m) and Rodna Mountains (ELA = 1697 m). The SE rising trend of the ELA corresponds well with the dominant palaeowind direction suggested in the Carpathian region and supports the prevalence of zonal circulation pattern in Central Eastern Europe during the culumination of the last glaciation.

Key words

glacier reconstruction • ELA • glacial sediments • Maramureș Mountains • Romania

Introduction

Glacial landforms in the Carpathians were already reported in the second half of the 19th century (Zejszner, 1856; Paul & Tietze, 1876; Tietze, 1878; Lehmann, 1881); however, regional reviews apart from an early synthesis by Pawłowski (1936) are largely lacking. Most of the research including well-dated geomorphological reconstructions of mountain glacier fluctuations was concentrated in the areas strongly shaped by glaciers such as the Tatra Mountains in the Western Carpathians (Engel et al., 2015; Makos et al., 2018; Kłapyta & Zasadni, 2018 and references therein) and the Retezat Mountains in the Southern Romanian Carpathians (Ruszkiczay-Rüdiger et al., 2016, 2021). Knowledge of the legacy of Pleistocene glaciation is still limited in many of the remaining mountain massifs in the Carpathians, particularly in low-altitude isolated mountain massifs where the glaciation was marginally developed.

The Eastern Carpathians are the north-easternmost mountain range in Central-Eastern Europe that bears witness of the Pleistocene glaciation (Fig. 1A). The presence of mountain glaciation was first described in the second half of the nineteenth century in Chornohora (Paul & Tietze, 1876) and Rodna Mountains (Lehman, 1891). Instead of an early recognition of glacial landforms in this area, the Eastern Carpathians are considered to be one of the least studied areas in terms of glacial geomorphology in Europe. Despite the documentation of glacial cirques (Mîndrescu & Evans, 2014; Mîndrescu, 2016), glacier reconstruction is hampered by the lack of mapped maximal moraines. One of such areas are the Maramureş Mountains, which represent the northernmost part of the Eastern Romanian Carpathians (Fig. 1A). Landforms of former glaciation are preserved here in several isolated mountain massifs that rise above 1800 m (sometimes only above 1700 m) in the north and 1850 m a.s.l. in the south (Mindrescu, 2016).

Mount Jupania (Crechela, 1853 m a.s.l.) is the south-easternmost glaciated area in the Maramureş Mountains. Due to its remote location, in close proximity to the state border between Romania and Ukraine, this area belongs to the most inaccessible natural regions in the Eastern Carpathians, hence its glacial landscape has been discovered relatively recently (Mîndrescu, 2001-2002). Two glacial cirques were identified on the north-eastern slope of the massif within the headwaters of the Ceremuşul Alb (ukr. Bilyj Cheremosh) valley (Mîndrescu, 2001-2002, 2016), while the extent of moraines was widely undetected due to inaccessible mountain terrain with dense forest vegetation. According to Urdea et al. (2011) the Jupania glacier was the largest (3.32 km^2) in the Maramureş Mountains, with a maximum length of 4.4 km, and a glacier front at 1205 m a.s.l. Recent large deforestation in the Jupania area opened the opportunity for detailed recognition of glacial landforms and deposits and reconstruction of former glaciation.

In this study we use a complex approach that includes geomorphological mapping and clast morphology analysis together with glacier surface reconstruction to provide the extent and equilibrium line altitude (ELA) of small palaeoglacier in the Mt. Jupania area. The objectives of this study are: (i) to document glacial landforms and deposits in study area; (ii) to reconstruct the extent of the glacier during the maximum ice extent (MIE); and (iii) to estimate the equilibrium line altitude during the MIE.

Study area

The Maramureş Mountains are a part of the Eastern Romanian Carpathians and stretch over 80 km along the border between Romania and Ukraine from the confluence of the Vişeu and Tysa rivers in the northwest to the Țibău valley in the south-east (Fig. 1A). The Maramureş Mountains are characterised by the presence of isolated mountain massifs reaching a maximum elevation of 1957 m at the top of Farcău Peak separated by the deeply incised (up to 1000 m) tributaries of the Vişeu

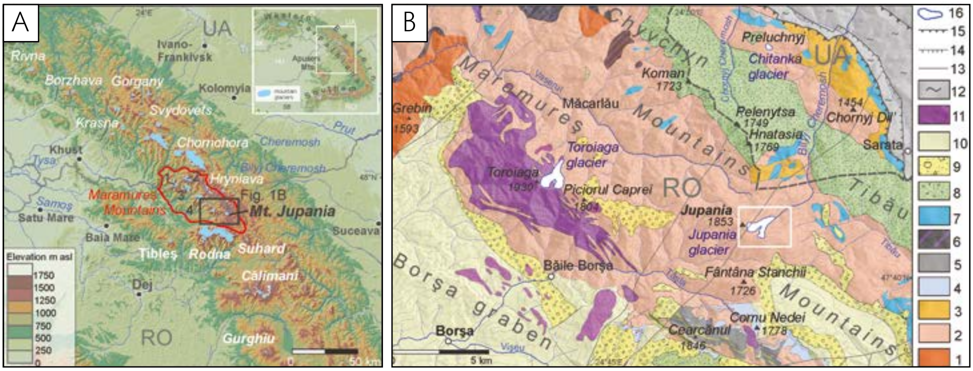

Figure 1. The study area. A – location of Maramureș Mountains and Mt. Jupania in the Eastern Carpathians (EC). All mountain ranges in EC with documented evidence of Pleistocene glaciation are marked in white Glaciated massifs in the western and central part of the Maramureș Mountains: 1 – Mt. Pop Ivan, 2 – Mezipotoki and Mt. Mica Mare/Nieneska, 3 – Mt. Farcău-Mihailecu, 4 – Mt. Pietrosu Bardaului. B – Geological map of the south-eastern part of the Maramureș Mountains (after: Harta geologica scara 1:200.000, foaia Viseu 1968, Institutul Geologic Bucuresti and Kräutner et al., (1983) with the marked extent of Pleistocene glaciers in Mt. Toroiaga, Jupania and Chitanka area (according to Kłapyta et al., 2023). B – location of the study area marked with white frame. Legend: Crystalline core (Bucovinian nappe stack) and its autochthonous sedimentary cover: 1 – gneisses, 2 – sericite and chlorite schists, 3 – mica schists and paragneisses, 4 – metatufites (Devonian), 5 – sericite and graphite schists, 6 – black shales and basalts (Jurassic), 7 – limestones and dolomites (Jurassic), 8 – Sojmul/Ceahlau conglomerates (Albian), 9 – Prislop conglomerates (Eocene), 10 – Post-orogenic Borșa flysch (Oligocene-Miocene); Neogene volcanics: 11 – andesites (Miocene); Outer Carpathians: 12 – flysch (undivided); 13 – faults, 14 – secondary thrusts, 15 – main thrusts, 16 – reconstructed glaciers

valley (Repedea and Ruscova). According to morphometric analysis (Mîndrescu, 2003), this area is divided into two parts: the larger northern Maramureș Mountains (Farcău 1957 m) extending over 243.3 km^2 and the much smaller southern Maramureș Mountains (Toroiaga 1930 m) with 123.7 km^2 of area. The main trait of this mountainous area is the presence of denudation surfaces and isolated peaks with the highest summits (Pop Ivan, Farcău-Mihailecu, Pietrosu Bardaului, Toroiaga, Cearcanu) that are found to the south from the main Carpathian drainage divide

The Maramureș Mountains are part of the Inner Eastern Carpathians built up by the Bucovinian nappe stack (Bucovinian and Subbucovinian nappes) which comprise the crystalline basement, the autochthonous sedimentary cover and intrusions of Neogene andesites (Kräutner et al., 1983; Săndulescu, 1984; Schmid et al., 2020) (Fig. 1B). During the Pleistocene the Maramureș Mountains were located between the much more strongly glaciated Chornohora Mountains to the north and the Rodna Mountains to the south (Fig. 1A). The traces of Pleistocene glaciation were identified in eight mountain massifs (Mt. Pop Ivan, Mezipotoki, Mica Mare/ Nieneska, Farcău-Mihailecu, Pietrosu Bardaului, Toroiaga, Jupania and Cearcănu; Fig. 1). In this study, we present a legacy of Pleistocene glaciation in the Jupania area which is the southernmost mountain area of the Maramureș Mountains (Fig. 1A).

Mt. Jupania is located at the main Carpathian drainage divide separating the basins of the rivers Prut and Cheremosh from that of the Tysa River. The NE slopes of Jupania are the source area of the Ceremușul Alb/ Bilyj Cheremosh River (61 km long, 606 km^2 drainage area), which flows further north on Ukrainian territory to the Prut river along the border between the historical regions of Bucovina and Galicia.

Due to its location close to the border between Romania and Ukraine, the region has been largely unaffected by deforestation until recently. Old growth spruce forest covered almost entirely the local catchments, reaching as high as 1600 m a.s.l. in Ceremuşul Alb/Bilyj Cheremosh River catchment and more than 1650 m a.s.l. on the western slope of Tâşla catchment. The remaining terrain up to the elevation of the main peak was covered by dense *Pinus mugo* shrubbery, thus Mt. Jupania was among the few Carpathian summits where all natural vegetation zones were kept intact and continuous and was regarded as an area of wilderness and high biodiversity. However, the last decade saw the massive deforestation of the largest part of Ceremuşul Alb catchment with ca. 500 ha of continuous old growth forest cleared between 2006 and 2009, which left behind only a narrow forest fringe at the upper part of the formerly forested area. The sudden, complete removal of forest vegetation contributed substantially to the increase in flood vulnerability in this area, particularly considering that extensive logging both inside and outside of Maramureş have already contributed to severe flood events in the past Knorn et al., 2012.

Methods

Geomorphological mapping

Geomorphological mapping of glacial landforms including lateral and terminal moraines, glacial cirques, and distribution of glacial deposits was conducted based on field investigations and a combination of available data sources: 1:25,000 topographic maps as well as by employing high-resolution satellite imagery (Google Earth and BingMaps) and the ALOS PALSAR (source: www.eorc.jaxa.jp) digital elevation model (DEM) with 25 m horizontal resolution. Glacial cirques and their floors were delineated using topographic maps and DEM-derived hillslope map and contours (Mîndrescu & Evans, 2014; Barr & Spagnolo, 2015). Cirque crests and floors were outlined on-screen in a GIS environment. The concave boundary between the cirque floor and headwall was drawn at a gradient of about 27° (Mîndrescu & Evans, 2014). Basic morphometric parameters of both cirque size (i.e., length, width, cirque/bottom area) and shape (maximum and minimum cirque slope) were measured using a 25 m DEM. Additionally, we estimated the maximum elevation of the cirque crest along with the minimum, maximum and mean elevation of the cirque floor. The difference between floor and crest elevation herein served as a measure of vertical cirque relief. The median cirque aspect was measured in an outward direction from the headwall to the threshold along a line dividing the cirque map area into two halves.

Clast morphology analysis

Analyses of clast morphology (clast shape and roundness) are commonly used in glaciated environments to distinguish between different erosional, transportation and depositional clast histories (Benn & Ballantyne, 1993, 1994; Glasser et al., 2009; Brook & Lukas, 2012). Clast morphology analysis (Lukas et al., 2013) was used in this study to confim the glacial origin of the mapped sediments. Moraine clasts commonly display a higher proportion of blade and elongate shapes and stronger rounding when compared to clasts of scree and gravitational mass movements sediments, but are much more angular compared to fluvial environments. Field sampling was undertaken at two sites including latero-frontal moraine (J-1 site) and ablation moraine (J-2 site) (Fig. 2). Measurements were confined to the sericite and chlorite schists that are the dominant type of clast lithology in the study area (Fig. 1B). Sampling sites were located along scarps of road cuttings. At each site the three orthogonal axes, a, b, c (long, intermediate, short) were measured for 50 clasts using a steel ruler. Clast roundness was determined visually for each clast on a modified Powers (1953) scale. Tertiary diagrams following Sneed and Folk (1958) were generated in the TRIPLOT Excel spreadsheet (Graham & Midgley, 2000). The C_{40} index (percentage of clasts

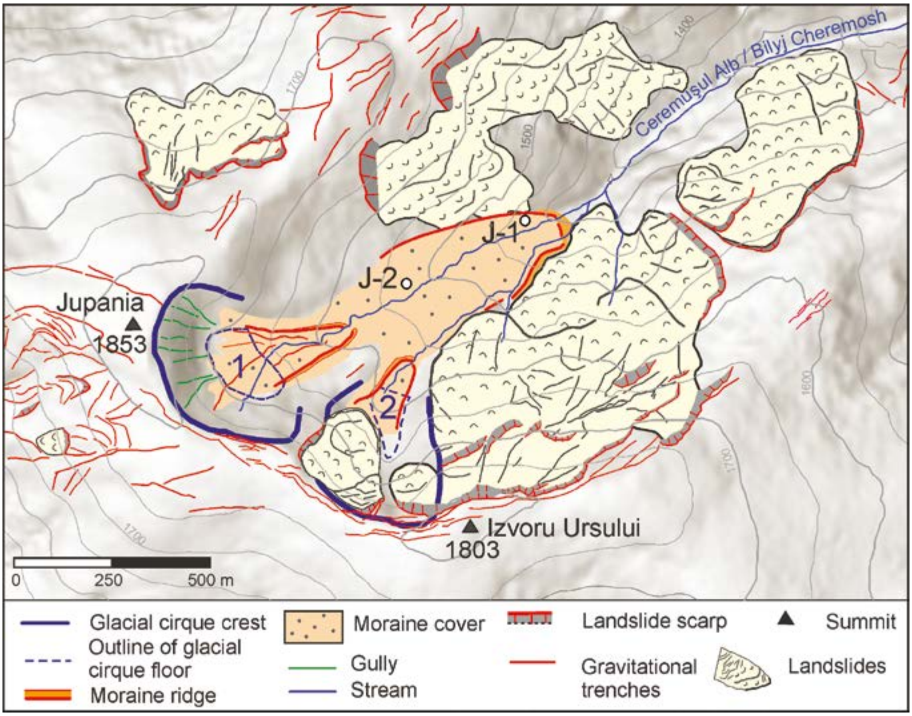

Figure 2. Geomorphological map of the Mt. Jupania and the headwaters of the Ceremușul Alb/Bilyj Cheremosh valley. Location of Jupania (1) and Izvoru Ursului (2) cirques. The clast-morphology measurement sites (J-1 and J-2) are marked

with c/a axial ratio ≤ 0.4) was subsequently calculated for each sample site. The RA ratio (the percentage of angular and very angular clasts in any sample) and RWR ratio (the percentage of rounded to well-rounded clasts; Lukas et al., 2013) were calculated for each site. In order to distinguish the transportation and depositional clast histories co-variance plots using both RA-C_{40} and RWR-C_{40} were presented (Benn & Ballantyne, 1994; Lukas et al., 2013) and compared with clast morphology data from the surrounding Rodna Mountains, where metamorphic clast lithology dominates (Kłapyta et al., 2021a, Supplementary Data 2).

Glacier reconstruction and ELA calculation

The shape and maximum extent of the Jupania glacier was reconstructed in the GIS environment based on the mapped distribution of lateral and frontal moraines and glacial deposits. Ice thickness was assessed along the flowline using a glacier profile model (Benn & Hulton, 2010) provided in the GlaRe, a semi-automated GIS-based method (Pellitero et al., 2015) with the use of basal shear stress values 50 to 100 kPa (Cuffey & Paterson, 2010) and a calculated valley shape factor. Ice thickness was adjusted to the mapped glacial landforms by fine-tuning their thickness using the basal shear stress values. The used basal shear stress values are typical for low gradient and relatively shallow ice thickness. We used a combination of manual and semi-automated methods (Zasadni & Kłapyta, 2014; Kłapyta et al., 2021a) to reproduce the glacier surface. The reconstructed glacier geometry has allowed for the calculation of the former ELAs using the area altitude balance ratio (AABR) method by incorporating both glacier hypsometry and variable glacier mass balance gradients in ELA calculations (Osmaston, 2005; Rea, 2009; Pellitero et al., 2015). We use the global median AABR value of 1.6 which is commonly used in paleoglaciological studies across the Carpathian and Balkan region (Ruszkiczay-Rüdiger et al., 2021; Zasadni et al., 2020; Kłapyta et al., 2021a,b) and close to global median value of monitored glaciers (1.56; Oien et al., 2021).

To allow comparison with other regional studies, we also report the median glacier elevation (accumulation area ratio; AAR 0.5), mean glacier elevation (AABR 1.0 or Kurowski's method; Braithwaite, 2015; Pellitero et al., 2015), and ELA at the size specific accumulation area ratio (ssAAR) (Kern & László, 2010). Due to the presence of gentle-sloping mountain-top surfaces which provide additional mass to the glacier surface by snowblow

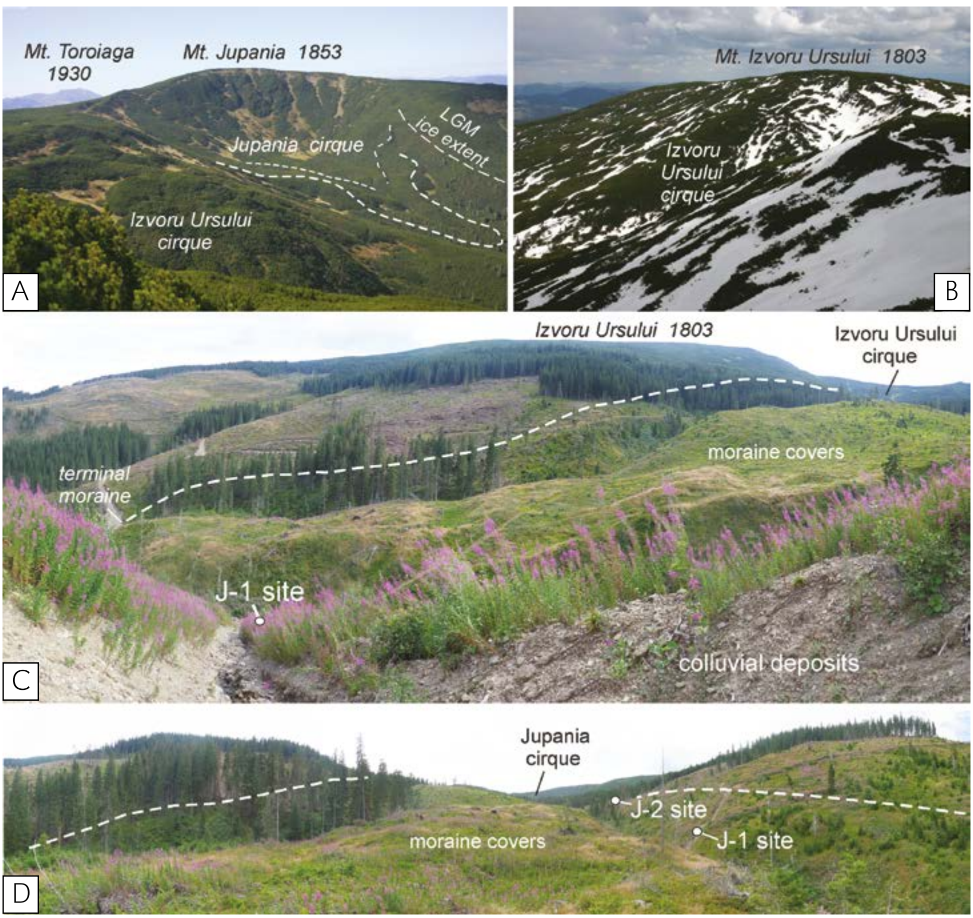

Figure 3. Geomorphological characteristics of the glacial landforms in Mt. Jupania area. A – The Jupania and Izvoru Ursului cirques, view from the Mt. Izvoru Ursului. Note the presence of recessional moraines on the Jupania cirque floor (dashed white line). B – View on the poorly developed Izvoru Ursului cirque. C – Well-preserved lateral and terminal moraines in the Ceremușul Alb/Bilyj Cheremosh valley. White dashed line marks the extent of glacial deposits. D – Moraine covers in the Ceremușul Alb/Bilyj Cheremosh valley, with the clast-morphology measurement sites J-1 and J-2; view up valley from the frontal moraine. White dashed line marks the extent of glacial deposits

we estimate a snow contribution area (sca) using the approach of Benn et al., (2005) which was also recently used in ELA calculation for the small paleoglaciers in the Eastern Carpathians (Kłapyta et al., 2022a,b). The 'sca' includes both the avalanche and snowblown areas. Avalanche areas were identified as the slopes surrounding the glaciers greater than 20° (Sissons & Sutherland, 1976). The snowblow areas were defined as terrain lying above the AABR 1.6 ELA which is laterally continuous to the reconstructed glacier and sloping toward its surface. In our calculations we include also uphill snowblow from gentle sloping ridge toward the glacier as a maximum slope angle of 10° away from the glacier (Coleman et al., 2009). Additionally, we calculate the avalanche factor (AF) defined as the ratio of the glacier area to the avalanche area (Sissons & Sutherland, 1976; Coleman et al., 2009), and the snowblow factor (SBF) expressed as the glacier area to the square root of the snowblowing area (Sissons, 1980; Mitchell, 1996). Finally, we calculate scaELA which includes inferred additional snow contribution area as maximum estimate of 'climatic' or regional ELA (Benn & Ballantyne, 2005).

Results

Glacial morphology

The extent and distribution of glacial landforms in the Jupania area are illustrated in Fig. 2, while examples of depositional landforms are shown in Fig. 3. Two glacial cirques with steep headwalls curved around an outsloping floor are the striking element of relief in the eastward oriented headwaters of the Ceremușul Alb valley (Fig. 2). The main cirque, Jupania, is a slightly open, definite cirque (grade 3) according to the criteria by Evans and Cox (1974), while Izvoru Ursului ranks as an open, poorly developed cirque (grade 4) (Tab. 1) with backwalls affected by rockslides (Fig. 2). Of the two cirques, Jupania (Fig. 2, no.1) contains a threshold, whereas Izvoru Ursului (Fig. 2, no. 2) does not show a threshold.

The cirque floors of the Jupania (~1660-1700 m a.s.l.) and Izvoru Ursului (~1580-1650 m a.s.l.) are among the highest cirques

Table 1. Basic morphometric parameters of glacial cirque size and shape in the Jupania area. Cirque numbers refer to Fig. 2

Morphometric characteristics	Jupania (1)	Izvoru Ursului (2)
Length [m]	434	523
Width [m]	615	462
Elongation L/W	0.71	1.13
Entire cirque area [ha]	25.9	22.7
Cirque floor area (ha)	6.47	2.61
Floor area/cirque area [%]	25.0	11.5
Perimeter [m]	1977	1799
Median aspect of cirque axis [°]	95	6
Maximum gradient [°]	48.6	40.8
Minimum gradient [°]	3.90	9.40
Maximal altitude of cirque crest [m a.s.l.]	1835	1806
Minimal altitude of cirque floor [m a.s.l.]	1651	1582
Mean altitude of cirque floor [m a.s.l.]	1680	1607
Vertical dimension [m]	184	224
Length/height range	2.36	2.33
Grade	3	4
Status of floor	drift	drift

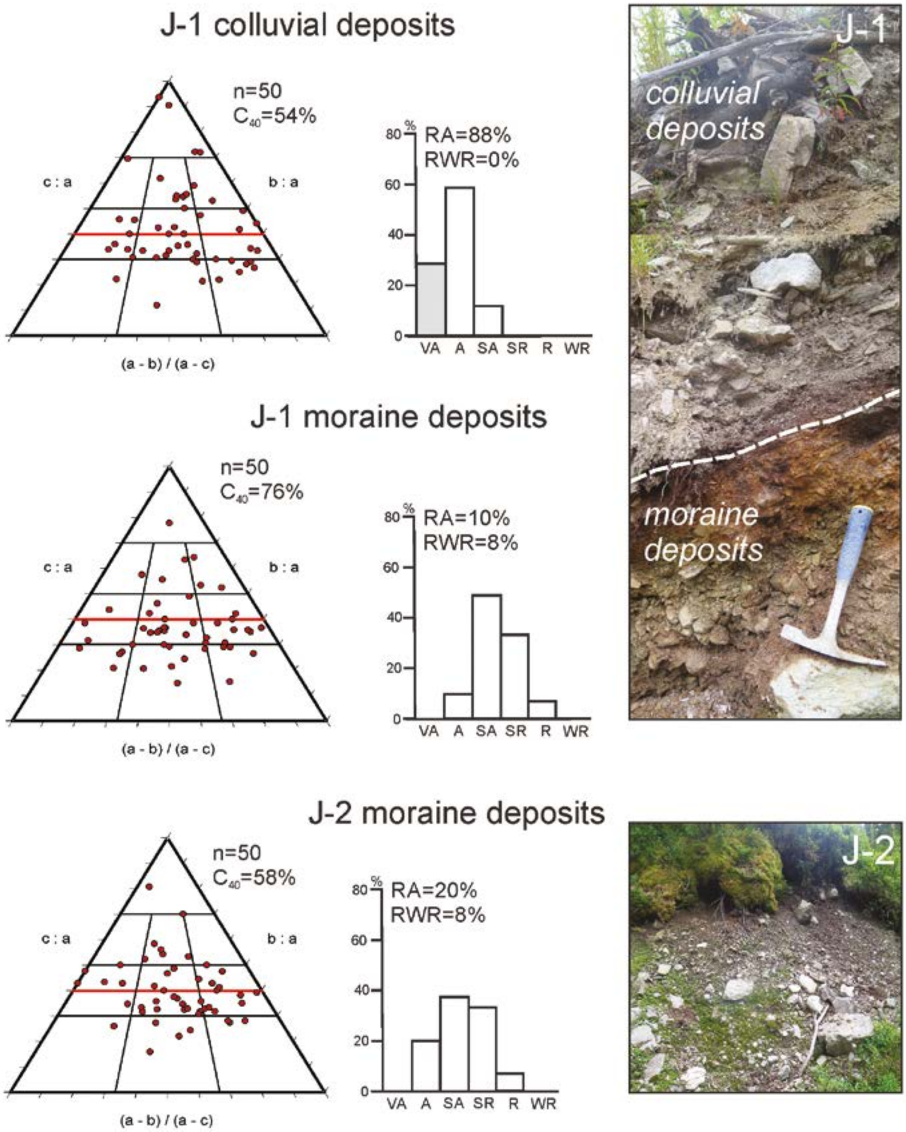

Figure 4. Clast shape data (ternary diagrams) and roundness (histograms) for sampled sites in the Mt. Jupania area. Red solid line marks the position of C_{40}. Site locations are depicted in Fig. 2. Abbreviations: n=number of sampled clasts; RA and C_{40} are defined in the text

in Maramureş Mountains (where cirque floor elevations range between 1400 and 1720 m a.s.l.). The headwall heights for the two cirques are in the range of 150 m (at 170 m, and 145 m, respectively) and reach their maximum altitudes between 1830 and 1850 m a.s.l. Based on their size (of around 20 ha), Jupania and Izvoru Ursului cirques fall into the category of average sized cirques for Maramureş Mountains (where the size ranges from 7 to 80 ha). Cirque width (of 600-700 m) is larger than cirque length (420-430 m) for both cirques, and exceeds the average width of cirques from Maramureş Mountains.

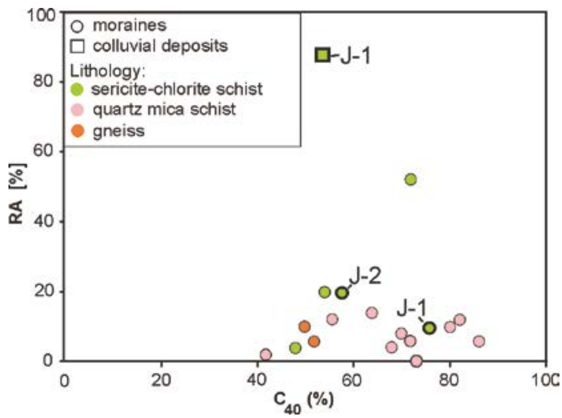

Figure 5. Co-variance plot for the moraine deposits in metamorphic lithology in the Mt. Jupania (bolded points) and the Rodna Mountains (Kłapyta et al., 2021a, Supp. data 2) using variations of RA versus C_{40}-index. Each plotted point represents a sample group of 50 clasts. Site locations in the study area are depicted in Fig. 2

For comparison, cirque width in Maramureș Mountains varies between 220 and 1200 m, while cirque length ranges between 300 and 800 m (Mîndrescu, 2016).

The MIE of the Jupania paleoglacier is marked by well-preserved lateral and terminal moraines descending down to the elevation of ~1400 m (Figs. 2, 3). Latero-frontal moraines reach up to 10-15 m high with the inner zone dominated by debris-rich hummocky topography which can be traced continuously from the terminal moraines to the cirque bottoms, typical for moraines of debris-covered glaciers. Latero-frontal recessional moraines are found in the elevation range 1570-1690 m and can be traced ~600 m up-valley from MIE moraines (Fig. 2). An outcrop next to a transverse path (J-1 site, Fig. 3) through the left-side lateral moraine and slope cover shows that moraine deposits of the MIE are locally overbuilt by 1-1.5 m thick colluvial cover resulting from the rock slope failure deposits coming from the valley sides above (Figs. 3, 4).

Clast-shape analysis

Ternary diagrams and roundness histograms for the sampled moraine sites are shown in Fig. 4, while the RA-C_{40} bivariate scatterplot is illustrated in Fig. 5. Sandy boulder-gravels, with minor cobbles and pebbles, dominate moraine deposits in the Mt. Jupania area. The petrographic composition of clasts is pure crystalline with a dominance of sericite and chlorite schists (Fig. 1A). Moraine deposits consist of diamict with a sandy-silty matrix supporting cobble- to boulder-sized clasts (Fig. 4). For the moraine deposits the C_{40} index range between 58% and 76%, indicating that moraines contain a high proportion of platy and blade-like clasts. A slightly lower value of C_{40} index (54%) was measured for colluvial deposits which covers the lateral moraine at J-1 site (Fig. 4).

RA values are generally low for the moraine deposits (10-20%) in comparison with colluvial deposits (88%) (Fig. 4), which points to the presence of more rounded clasts in moraine sediments compared to slope deposits. This is also well-expressed in RWR values, which are higher (8%) for moraines than colluvial deposits (0%) (Fig. 4). The co-variant plot (Fig. 5) demonstrates a narrow distribution of RA/C_{40} for moraine deposits further suggesting active glacial transport (lodgement till) of debris and a passive transport of clasts within slope deposits.

Glacier reconstruction and ELA calculation

During the MIE the cirque-valley glacier (0.85 km^2 area, 1.7 km long) developed on the NE slope of Mt. Jupania (Fig. 6). The area-averaged glacier thickness was 20 m, whereas the maximum thickness reached 43 m (Fig. 6). The ELA (AABR 1.6) of the reconstructed Jupania glacier was at 1630 m a.s.l., which is slightly higher compared to the value obtained by applying the AAR 0.5 method (1609 m a.s.l.) and 34 m above the value determined using the ssAAR method (1596 m a.s.l.; AAR 0.47). The total snow contribution area ('sca') in the study area was 0.82 km^2 which is similar to area of reconstructed glacier therefore the snowblow factor (SBF) was close to one (0.98). This suggest that additional snowblow delivery from mountain-top surfaces was important in glacier mass balance and thus in the size of glaciers and the ELA position. The snow contribution by avalanches was less important in total sca (avalanche factor, AF = 0.34). The inclusion of 'sca' in the ELA calculation (Benn & Ballantyne, 2005; Kłapyta et al., 2022a, b) gives scaELA at 1676 m which is 46 m higher than the AABR 1.6 ELA (Fig. 6).

Discussion

A precise understanding of past glacier evolution requires reliable spatial data on glacier's extent and ice-surface geometries. This requires knowledge on the position of ice marginal landforms, which serve as major features in the reconstruction of former glacier geometry (Barr & Lovell, 2014). With the availability of medium-resolution global digital surface models, mapping of large-scale erosional traces of glaciation as glacial cirques, is now possible better than before (Lopes et al., 2018; Evans et al., 2021). By contrast, the effective mapping of terminal moraines is not easy using remote-sensing data alone, especially in densely vegetated mountain areas such as the Carpathians where ice-marginal moraines could be erroneously regarded as landslides or other colluvial-type deposits (Sawicki, 1912; Pawłowski, 1936; Świderski, 1938). Therefore, careful field examination is required for both the geomorphological and sedimentological context of moraines before they are used to reconstruct former glacier geometry and infer past climatic conditions (Zasadni et al., 2021).

The legacy of former glaciation in the Eastern Romanian Carpathians consists of a vast array of landforms and deposits that can be used to reconstruct records of former climatic and landscape changes. Nevertheless, because of restricted mapping of glacial landforms and deposits our knowledge of the glacial history of the Eastern Carpathians is limited (Pawłowski, 1936). Recent research efforts (Kłapyta et al., 2021a,b, 2022a,b) based on the same methodological approach of ice surface and glacier ELA reconstruction

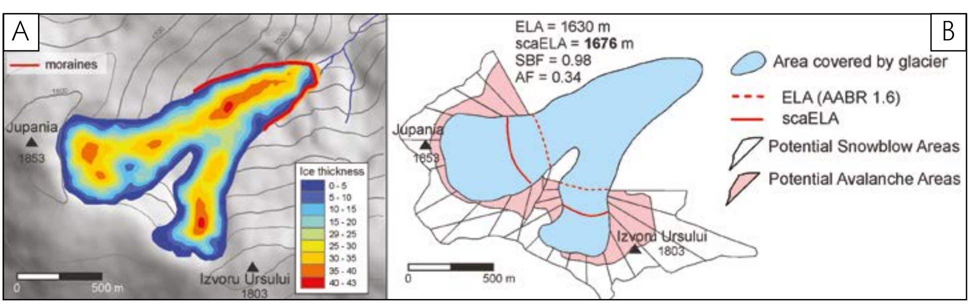

Figure 6. Glacier and equilibrium line altitude reconstruction of the LLGM advance in the Jupania massif A – Glacier thickness of the Jupania palaeoglacier. B – Distribution of potential snowblow areas and potential avalanche areas of the Jupania palaeoglacier. The position of ELA (AABR 1.6) and scaELA are marked in dashed and solid red lines, respectively

refined our knowledge of the climatic character of glacial environments in several mountain massifs in the Eastern Carpathians during local MIE which is assumed to coincide with the global LGM (Kłapyta et al., 2021a, b). However, in some low-altitude isolated mountain massifs the presence of former glaciation is still under debate (Sîrcu, 1963; Mac et al., 1990; Urdea et al., 2011) mostly due to the lack of documentation of moraine landforms and deposits.

In the Jupania area fresh-shaped MIE moraines are preserved at ~1400 m. In the valley below the terminal moraines no large boulders or patches of till deposits have been found, which could indicate erosion of older glacial sediments. In the Eastern Carpathians strongly degraded moraine deposits of pre-LGM glaciations were documented locally ~1 km beyond the fresh-shaped moraines in the largest glaciated valley systems in the Chornohora, Svydovets (Świderski, 1938; Kłapyta et al., 2021b) and Rodna Mountains (Gheorghiu, 2012; Kłapyta et al., 2021a). In the neighbouring Rodna Mountains the lowest moraines in Pietroasa and Cimpoieşului valleys were found much lower (840-850 m a.s.l.) than in the study area, however based on its significant surface degradation and relative sediment preservation they represent older pre-LGM moraines (Şesura unit, likely MIS 6) (Kłapyta et al., 2021a). It should be emphasized that these relatively low elevation moraines is only local phenomenon controlled by enhanced topography (exceptionally steep valley profiles), which enable glaciers to expand much further beyond the LGM position than in other Rodna valleys during pre-LGM glaciations. On the strongly glaciated northern slope of Rodna Mts, the lower limit of the LGM moraines (Pietroasa unit) typically occurs between 1100-1150 m a.s.l. and the mean ELA was at 1657 m a.s.l. (Kłapyta et al., 2021a). The discussion about the age of the MIE moraines in the Rodna Mountains, which age was proposed with ^{10}Be exposure age dating between 37.2-26.6 ka (Gheorghiu, 2012) is yet to be concluded. The geomorphological analysis and results of weathering grade studies (Kłapyta et al., 2021a) indicate that ^{10}Be dated MIE moraines in the Rodna Mountains represent strongly degraded Şesura unit moraines (pre-LGM) where due to prolonged time of boulder exhumation exposure age of boulders may be substantially younger than the moraine. The oldest exposure age of boulders on fresh-shaped Pietroasa moraines (17.5±1.6 ka; Gheorghiu, 2012) support moraine deposition that falls within the global LGM time range (Kłapyta et al., 2021a).

The bedrock lithology of the Jupania area is characterised by rather soft and highly anisotropic metamorphic rocks (sericite and chlorite schists) (Fig. 1B). The initial clast morphology features high clast platiness and high angularity which are well expressed in the high C_{40} (80%), high RA (80%) and low RWR values (0%) found in colluvial covers (Fig. 4). Our sedimentological data indicates that pronounced clast reworking due to glacial transport was effective enough to remove the primary angularity of metamorphic schists clasts and its progressive rounding. This is well expressed in the lower RA (10-20%) and higher RWR (8%) values obtained for moraine sites located 0.7-1.0 km down valley from cirque headwalls (Fig. 4). A reduction of clast angularity without substantial reduction of clast platiness was also found in the most extensive moraines in the Rodna Mountains found 4-5 km from cirque headwalls (Kłapyta et al., 2021a, Supp data 2). These results correspond well with the finding of Lukas et al. (2013) that for soft and highly anisotropic metamorphic rocks, roundness is the dominant discriminator of glacial reworking.

The Maramureş Mountains are typical for asymmetrically glaciated ranges that can support glaciers only in the most favourable aspects (Evans, 1977; Evans & Cox, 2005). The majority of cirques are NE-oriented (0-90 degrees), including Jupania (95°) and Izvoru Ursului (6°), which range at the extremities of the quadrant. Only marginal cirques are NW-oriented in this mountain range (264-355°). The Jupania palaeoglacier developed on the N and NE, leeward slopes where snow drift accumulation and avalanches from collapsing cornices were important controls for the development of marginal glaciation. The development of glaciation on poleward (N) and leeward (NE) slopes highlights the effect of solar radiation and wind effects on glacier ablation and accumulation (Evans, 2021).

The presented results from the Maramureş Mountains shed new light on the glacial evolution of low-altitude isolated mountain massifs located between the higher and more strongly glaciated Chornohora (2061 m) and Rodna Mountains (2303 m). The traces of Pleistocene glaciation were identified in eight mountain massifs in the Maramureş Mountains: Mount Pop Ivan (Sawicki, 1911 Hnatiuk, 1987; Mac et al., 1990; Kravchuk, 2021), Mezipotoki (Mindrescu, 2016), Mica Mare/Nieneska (Kondracki, 1935, 1937), Farcău-Mihailecu (Sawicki, 1911; Sârcu, 1963; Mac et al., 1990; Mîndrescu, 1997), Pietrosu Bardăului (Sawicki, 1911; Sîrcu, 1963), Toroiaga (Sawicki, 1911) Jupania (Mîndrescu, 2002) and Cearcănu (Mindrescu, 2016). According to Urdea et al. (2011) the Jupania glacier was the largest (3.32 km^2) in the Maramureş Mountains, with a maximum length of 4.4 km, and a glacier front at 1205 m a.s.l. According to our results the presence of glacier of such size in this area, is unlikely. Our results show that the Jupania palaeoglacier was three times smaller with 0.85 km^2 area, and 1.7 km long. According to Kłapyta et al. (2023), the largest palaeoglacier documented in the Maramureş Mountains was only 1.68 km^2 area, 3.16 km long (the Rugaşu valley glacier). The Rugaşu glacier was located on the southern slope of the Farcău- Mihailecu massif in the north-western and the highest (1957 m a.s.l.) part of the Maramureş Mountains (Fig. 1). This part of the mountains is also characterised by the presence of the largest cirques, such as Vârtop (81 ha) in Farcău-Mihailecu massif, Dezeskul Grun (58 ha) in Pop Ivan massif, and Bardău (54 ha) in Pietrosu Bardăului area (Mindrescu, 2016). In this context the Jupania palaeoglacier was located in the part of the mountains where development of glaciation was less favoured. This is indicated by our ELA estimation (1630-1676 m) which fits well to the general SE rising ELA trend recognized along the Eastern Carpathians with a mean rise of ~3 m·km^{-1} (Kłapyta et al., 2021a, b; 2023).

Assuming that MIE in the study area is attributed to the Last Glacial Maximum (LGM) the LGM ELA in the eastern Maramureş Mountains is 114-160 m higher compared to the Chornohora Mountains (1516 m; Kłapyta et al., 2021 b) and 20-67 m lower than in the Rodna Mountains (1697 m; Kłapyta et al., 2021a). The SE trend of the ELA increase in the Eastern Carpathians is consistent also with the southeastward rise of cirque floors (Kłapyta et al., 2021b) and reflects the prevailing zonal circulation pattern in Central Eastern Europe with dominant north-westerly and westerly palaeo-wind direction (Kłapyta et al., 2021a,b, 2022a,b, 2023). Therefore, the southern Maramureş Mountains was climatically less favoured area for ice accumulation due to precipitation starvation caused by orographic shading in the mountain interior.

Conclusions

Based on the field documentation of glacial landforms and deposits we established the limits of maximal ice extent in the headwaters of the Ceremuşul Alb/Bilyj Cheremosh valley during the Late Pleistocene. The study area stands out as an isolated and poorly accessible part of the Maramureş Mountains where the legacy of Pleistocene glaciation has only recently been discovered. Our data confirm usefulness of clast morphology analysis in distinguishing between glacial and nonglacial sediments in marginally glaciated area. The glacial landscape is indicative of the former presence of small cirque-glacier, the Jupania palaeoglacier. It covered an area of 0.85 km^2 and was 1.7 km long, which is three times smaller than the previous assumptions. The estimated equilibrium line altitude (AABR ELA 1.6) was placed at 1630 m a.s.l., however, when taking account additional snow-blow delivery from gentle-sloping mountain-top surfaces, the resulted ELA is placed even higher at 1676 m. The reconstructed ELA in the south-eastern part of the Maramureș Mountains fits well with the south-east rising ELA trend documented between the Chornohora Mountains and the Rodna Mountains, and reflects a dominant NW precipitation–wind regime in the Eastern Carpathians during the maximal ice extent.

Editors' note:
Unless otherwise stated, the sources of tables and figures are the authors', on the basis of their own research.

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