Quaternary International 388 (2015) 156–167

Contents lists available at ScienceDirect

Quaternary International

journal homepage: www.elsevier.com/locate/quaint

Deglaciation constraints in the Parâng Mountains, Southern Romania, using surface exposure dating

D.M. Gheorghiu [a,*], M. Hosu [b], C. Corpade [b], S. Xu [c]

[a] CIAF- NERC, Scottish Universities Environmental Research Centre, East Kilbride G75 0QF, UK
[b] Faculty of Geography, Babeş-Bolyai University, Str.Clinicilor Nr. 5-7, 400006, Romania
[c] AMS, Scottish Universities Environmental Research Centre, East Kilbride G75 0QF, UK

ARTICLE INFO

Article history:
Available online 11 June 2015

Keywords:
Glaciers
Glacial geomorphology
Deglaciation
Surface exposure ages
Parâng Mountains
Romania

ABSTRACT

Cosmogenic nuclide surface exposure ages have been widely used to constrain glacial chronologies in the European regions. This paper brings new evidence that the Romanian Carpathians sheltered mountain glaciers in their upper valleys and cirques until the end of the last glaciation. Twenty-four ^{10}Be surface exposure ages were obtained from boulders on moraine crests in the central area of the Parâng Mountains, Southern Carpathians. Exposure ages were used to constrain the timing of the deglaciation events during the Late Glacial. The lowest boulders yielded an age of 13.0 ± 1.1 (1766 m) and final deglaciation occurred at 10.2 ± 0.9 ka (2055 m). Timing of the Late Glacial events and complete deglaciation reported in this study are consistent with, and confirm, previously reported ages of deglaciation within the Carpathian and surrounding European region.

1. Introduction

During the Last Glacial Maximum (26–21 ka; Peltier and Fairbanks, 2006), the southern and eastern part of the continent was affected by mountain glaciation or small ice caps, specifically in south Germany, the Pyrenees, the Alps, the Vosges and Jura Mountains, the Carpathians and the Ural mountains (Ehlers et al., 2011). Several smaller ice caps formed in the Massif Central, Vosges and Jura Mountains (Gillespie and Molnar, 1995). The Alpine ice cap drained through several ice streams that occupied the main valleys, forming the largest glacial system beyond the southern limits of the massive European ice sheet (Florineth and Schlüchter, 1998; Ivy-Ochs et al., 2008). The glaciers in the eastern Alps attained their maximum position as piedmont glaciers between 24 and 21 ka, rapidly decaying afterwards with occasional short time ice oscillations and stillstands (~16 ka), followed by other re-advances during the Late Glacial (van Husen, 1997; Reitner, 2007). Glaciers in the Tatra Mountains steadily retreated from the last maximum ice advance after 21.5 ka and two ice readvances occurred before the final deglaciation at 9.5 ka (Makos et al., 2013). Deposition of moraines in the Ukrainean Carpathians occurred during the last glacial episode at ~12 ka (Rinterknecht et al., 2012). South-eastern Europe was characterized only by restricted mountain glaciation with various records for the last glaciation (e.g. Marjanac and Marjanac, 2004; Woodward et al., 2004; Hughes et al., 2006; Hughes and Braithwaite, 2008; Milivojevič et al., 2008; Kuhlemann et al., 2009; Hughes et al., 2010). In Turkey, ice reached maximum extent before 26.1 ka (Akçar et al., 2007; Sarıkaya et al., 2009) and final deglaciation is recorded in southern Turkey at ca. 11.5 ka (Zahno et al., 2010).

The Romanian Carpathians have been mapped for over a century (e.g. Lehmann, 1881, 1903; Pawłowski, 1936; de Martonne, 1907; Sârcu, 1963; Morariu, 1981). Based on glacial geomorphology, initial assumptions were that two distinct glacial stages occurred during the Pleistocene, separated by an interglacial period, in addition to other smaller readvances or recessional stages (Posea, 1974). The recent mapping studies suggested that the Quaternary glaciations in the Romanian Carpathians were more extensive than previously thought (Urdea and Reuther, 2009; Urdea et al., 2011). According to temporal constraints using ^{10}Be ages, Gheorghiu (2012) suggest that ice during the maximum extent was more extensive and reached lower than 700 m altitude in north Romania. Three glacial stages were identified in a complex study in the Retezat Mountains, to the west of the Parâng Mountains (Reuther et al., 2007). Here, the maximum extent was not dated numerically, however, surface exposure dating indicate that deglaciation during the Late Glacial

* Corresponding author.
E-mail address: Delia.Gheorghiu@glasgow.ac.uk (D.M. Gheorghiu).

http://dx.doi.org/10.1016/j.quaint.2015.04.059

occurred at 16.1 ± 1.7 ka, with the final ice melt at 13.6 ± 1.5 ka and 11.4 ± 1.3 ka (Reuther et al., 2007). North of the Parâng Mountains, the relative chronology of glacial events was established based on the frontal moraines found in the Şureanu Mountains (Urdea and Drăguţ, 2000; Drăguţ, 2004). A more recent study in the Făgăraş Mountains has revealed that the ice retreat occurred between 17.4 ± 3.1 ka and 12.6 ± 2.0 ka (Kuhlemann et al., 2013a).

Several studies in the Parâng Mountains have focused on the morphometric aspects of the glacial cirques and their comparison with other Carpathian regions (Grozescu, 1920; Lişteveanu, 1942; Sârcu and Sficlea, 1956; Iancu, 1958, 1963, 1972; Vuia, 2002, 2003). Some erratic boulders were identified on the Lotru valley at an elevation of 1550 m (Grozescu, 1920), but maximum extent of ice was considered to be at 1340 m where the rest of a frontal moraine was found (Lişteveanu, 1942). A single surface exposure age is stated for the deglaciation timing in the Parâng Mountains at 17.9 ± 1.6 ka, however, no details about sample and location of the study is given (Urdea and Reuther, 2009). No other known study has focused on the numerical dating of the deposits in the glaciated areas of the Parâng Mountains.

The present paper utilizes the cosmogenic nuclide surface exposure dating to constrain the deglaciation in the alpine area of the Parâng Mountains, located in the Southern Romanian Carpathians (Fig. 1). The rapid changes in the North Atlantic region are well documented in western and central European regions (Peltier et al., 2006). However, areas located further away from the Atlantic do not have enough proxy records for the Late Glacial period. No doubt more information would increase the knowledge on the pattern and dynamics of the atmospheric circulation across Europe at the end of the last glaciation (Würm). Although some paleoglacial reconstructions have been conducted in the Parâng Mountains (Iancu, 1972; Vuia, 2003), this area has not benefitted from timing constraints using numerical dating. Thus, the current study is an additional contribution to the systematic investigation of key sites in the Romanian Carpathians, which aims to establish a complete history of the palaeoenvironment since the last ice maximum extent.

The main objectives of this paper are 1) to temporally constrain the deglaciation events during the Late Glacial using in-situ ^{10}Be surface exposure dating on the remnants of a late Pleistocene glacial complex in the central Parâng Mountains and 2) to provide an overview of the palaeoenvironment in the Romanian Carpathians at the end of the last deglaciation.

2. Regional and climatic setting

The Parâng Mountains are part of the highest mountain range of Romania located in the Southern Carpathians (Fig. 1). The area is bordered by the Şureanu Mountains to the north and Getic Subcarpathians to the south. The Retezat Mountains are located to the west of the Parâng Mountains and is separated from the Făgăraş Mountains to the east by the Olt valley. The main ridge is mostly located above 2000 m (>30 km in length) and is sinuously orientated from west to east.

Geologically, the Parâng Mountains consists mostly of crystalline rocks originated in the pre-Alpine orogenic events (Bercia et al., 1967; Savu et al., 1968). They are mainly represented by Proterozoic and Paleozoic metaclastic suite, metamorphosed up to amphibolite facies, the dominant rock types being quartzitic gneisses and micaceous schists. Large granitic bodies intruded the metasedimentary pile, their elongated shape resulting in the ridge orientation. The sedimentary cover is preserved only in patches, including Upper Paleozoic and Mesozoic massive limestones and

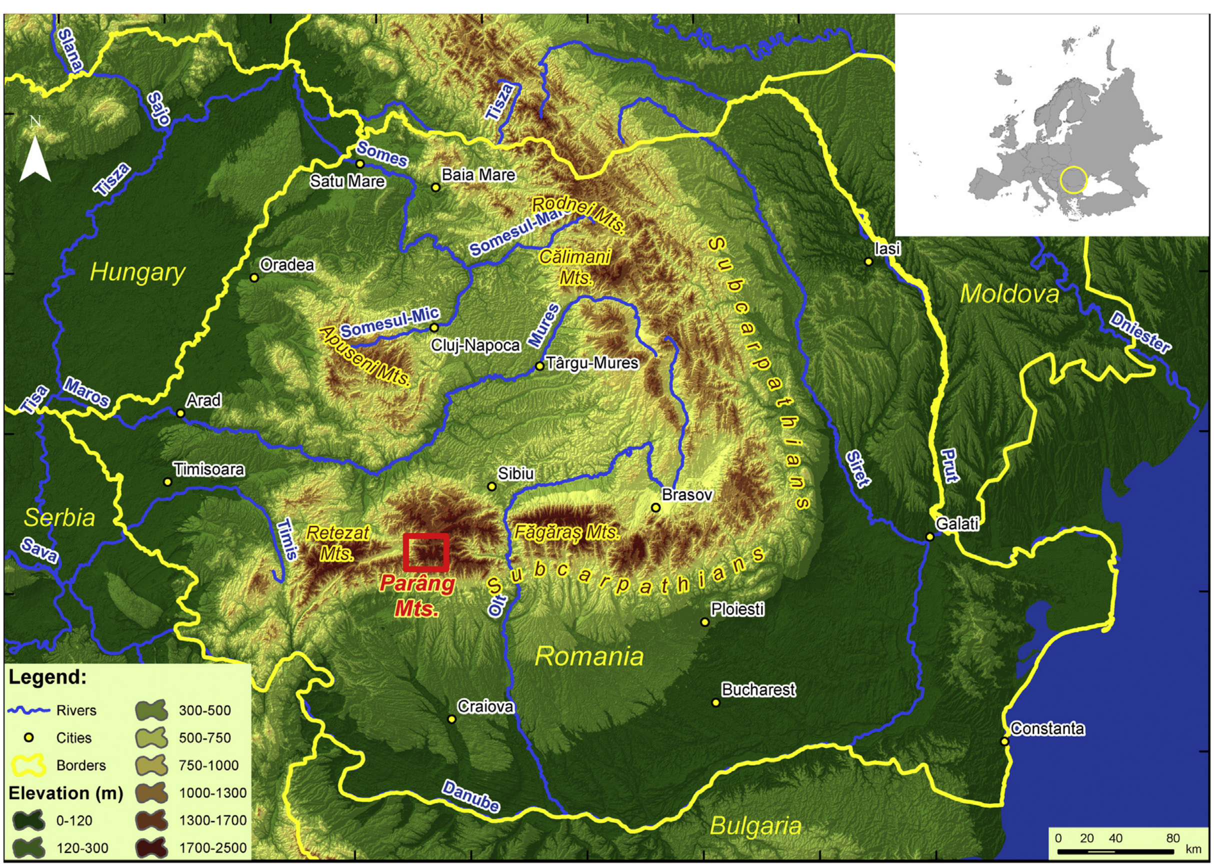

Fig. 1. Parâng Mountains. Inset map shows location of the study area in Europe.

conglomerates and Cenozoic rocks. Jurasic limestone areas are mainly found on the southern sector of the mountains (Bercia et al., 1967; Savu et al., 1968).

The Southern Carpathians are characterised by a continental temperate climate. The dominant wind directions are west and south-west. In case of southern or northern winds, the west-east orientation of the mountain ridge acts as a barrier between the humid Mediteranean air masses and the colder air masses from north. Mean annual temperatures in the Parâng Mountains are below 0 °C in the high alpine areas (>2000 m). Annual precipitation ranges between 800 mm on the lower ground and >1200 mm in the higher areas. Frequent solid precipitation occurs in the higher areas above 1500 m altitude, but the lower areas are affected by foehn (Posea, 2003, 2004; AMN, 2008).

The area referred to in this study (~13 km^2) is located in the central part of the Parâng Mountains (Figs. 1 and 2). It includes the five cirques of North Mohoru, Zănoaga Pietroasă, Gâlcescu, Căldarea Dracului and Zănoaga Mare, all draining in the Lotru valley (Fig. 2).

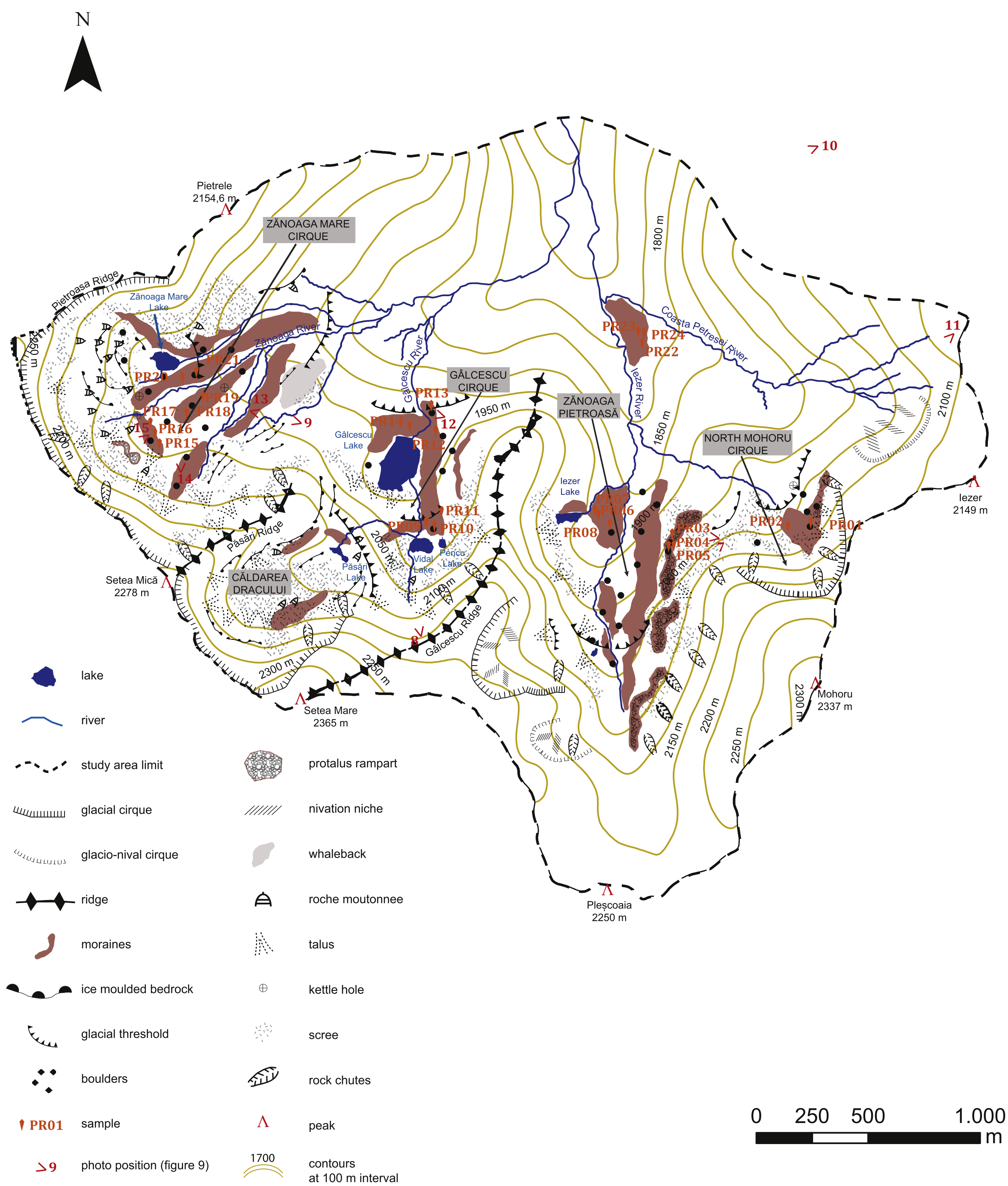

Fig. 2. Geomorphological map of the central part of the Parâng Mountains.

3. Materials and methods

The field area extends between 2365 m and 1400 m altitude. We used topographic maps (1:50000; 1:25000), Google Earth high-resolution imagery and orthophotmaps (50 cm resolution) to identify glacial features on the north-eastern slopes of source area of the Lotru valley (Fig. 2). Glacial landforms were difficult to map because of the heavily forested area at lower elevations in the Lotru valley. The glacial landforms were initially mapped and sampled after field observations. We collected 24 samples for ^{10}Be dating in quartz from boulders on moraines in 4 cirques (North Mohoru – 2 samples; Zănoaga Pietroasă – 6 samples; Gâlcescu – 6 samples; Zănoaga Mare – 7 samples) and in the upper part of the Lotru Valley (3 samples) (Table 1). The position, altitude, topographic shielding, dimension of boulders were recorded in the field (Table 1). Maximum 3 cm were removed from the boulder upper surfaces which were carefully selected to minimize any potential impact of rolling from cirque walls, overturning or shielding. Samples were taken from large (>1–2 m a-xis) gneiss boulders well embedded in the ground (Table 1). A GPS unit (Garmin Oregon 450) was used to record sample locations and elevations. Topographic shielding factor was determined for all the surrounding mountain slopes at each of the sampled localities (Dunne et al., 1999).

All samples were processed at the CIAF - SUERC (Cosmogenic Isotope Analysis Facility - Scottish Universities Environmental Research Centre), using procedures based on Kohl and Nishiizumi (1992) and Child (2000). Details of sample locations and relevant analytical data are given in Table 1. Ratios of the radionuclide to the stable nuclide were measured at the SUERC AMS Laboratory using the 5 MV accelerator mass spectrometer (Xu et al., 2010). Measured ^{10}Be/^{9}Be ratios were corrected by full chemistry procedural blanks (2–5 % of the sample ratios). The ^{10}Be ages were calculated with the Cronus Earth online calculator v. 2.2 (http://hess.ess.washington.edu/; Balco et al., 2008) using a ^{10}Be half-life of 1.36 Ma and the SLHL production rate of 4.39 ± 0.37 atoms g^{-1}–a^{-1} (Lm scaling) obtained from age-constrained calibration measurements (Balco et al., 2008). The calculated age uncertainties are expressed as ±1σ (Table 1). The minimum exposure ages are calculated assuming zero erosion rates for the sampled surfaces since deposition. If erosion was considered, the 3 cm of micro-relief found on quartz veins on several sampled surfaces would increase our ages only with ~3% for an 11 ka exposure.

4. Geomorphological setting

The results of our geomorphological mapping in the central Parâng Mountains is shown on Fig. 2. Based on the morphologic and morphometric characteristics, there are two types of cirques in this upper part of the Lotru Valley: cirques with quasi-horizontal floor located at 1950–2050 m towards the higher end of the valleys, likely pre-glacial, and smaller hanging cirques on the side walls at 2100–2200 m altitude.

4.1. Zănoaga Pietroasă cirque

The Zănoaga Pietroasă cirque is located in the easternmost part of our study area, on a south-north direction, and is drained by the Iezer rRver (Fig. 2). In spite of a large accumulation area, there is no defined cirque at the higher end of the valley. Here, the river has been eroding headwards, most likely post-glacially. The side walls of the Zănoaga Pietroasă cirque are generally steep, and frost-shattered material has rolled down to the bottom of the walls, a process which is still active today.

The western side wall of this glaciated area shelters a smaller glacio-nival cirque in the south, with a very asymmetric transverse profile. It is a highly degraded area due to active ravines throughout. To the north, a small hanging cirque is an indication of a former accumulation area that fed into the main glacier of the Iezer valley (Fig. 2). The cirque's floor is located at 2030–2060 m altitude, ~40 m above the main valley floor. Evidence of glacial erosion is indicated by a reverse slope on the floor of this cirque, covered with frost-shattered material (Fig. 3a). The rest of the western wall is very steep (up to 200 m in height) and cut by a deep rock chute towards the north (Figs. 2 and 3a).

The eastern side wall indicates very active slope movement processes, with numerous talus cones at the bottom. The most highly developed cirque in the Iezer valley is located at 1960–1980 on the northern side of the Mohoru peak (2336 m), here named the North Mohoru cirque. This glacial cirque has a typical amphitheatre shape with a north-eastwards direction. An arcuate moraine is located towards the back wall and has large boulders (>1–2 m a-axis) well-embedded in its surface (Figs. 2 and 3a).

The main Iezer valley is covered in a series of glacial deposits on a 900 m long and 120 m wide surface which the river has mildly incised since glacial retreat. Several micro depressions (<3 m in diameter) can be found in these deposits, suggesting in situ melting of ice during deglaciation.

The most notable glacial landform is located along the eastern valley side (~600 m long) (Fig. 2). This landform was probably deposited as a lateral moraine, with very steep ice-proximal sides and mild ice-distal slopes. It consists of blocks of rocks of various dimensions (from few cm to few m). The top of the moraine has been extensively modified into several arched features through the continuous addition of frost-shattered material from the side wall, especially towards the southern part where valley side walls are closer.

Further down valley, a clear reverse slope has been formed through intense glacial erosion, just before the glacial threshold that separates the Iezer upper valley from its lower part towards the Lotru valley. Here, a moraine was deposited allowing the formation of the Iezer Lake (>10 m in diameter) at 1900 m altitude. Material transported by the river postglacially has filled more than half of the lake surface and its former extent can be easily traced. The moraine has been deeply incised by the Iezer river during the Holocene.

Above the confluence of the Coasta Pietroasei and the Iezer rivers there is a lateral moraine deposit ~4–5 m above the valley floor. The deposit dips ~15° northwards and is covered in semi-rounded and semi-angular boulders of variable dimensions (up to 4 m a-axis).

4.2. Gâlcescu cirque complex

The Gâlcescu cirque complex is located in the western part of our study area, in a south-west to north-east orientation (Figs. 2, 3b–d). To the east, it is separated from the Iezer valley by the Gâlcescu Ridge. The western limit is made by the Pietroasa Ridge, a secondary ridge extending from the main ridge (2250 m) in the south towards Pietrele peak (2154 m) in the north.

The Gâlcescu cirque complex consists of three individual cirques: Gâlcescu, Căldarea Dracului and Zănoaga Mare (Fig. 2). Glacial erosion and deposition processes left a very detailed glacial imprint in all these cirques. The cirques are located on two different levels. The lower cirques (1950–2050 m) are very well-developed and consist of 2–3 reverse bed slopes in which lakes have formed. The higher cirques (2100–2200 m) are smaller and with a typical amphitheatre shape.

The Căldarea Dracului is a hanging cirque located below the Setea Mare (2365 m) and Setea Mica (2278 m) peaks and it is separated from the Zănoaga Mare cirque by the glacial transfluence

Table 1
Sample characteristics, ^{10}Be concentrations and calculated surface exposure ages.

Lab ID	Lithology	Sampled surface	Size of sampled surface (m; L × w × h)	Latitude (°N)	Longitude (°E)	Elevation (m)	Shielding[a] (factor)	Thickness (cm)	Quartz mass (g)	[^{10}Be][b] (x10^4 atom g^{-1})	Exposure age[c] (ka)
Iezer valley											
PR01	Gneiss	Boulder	5.0 × 3.0 × 2.2	45.34	23.63	2034	0.9477	3	10.99	24.34 ± 1.08	11.2 ± 1.1 (0.5)
PR02	Gneiss	Boulder	1.0 × 1.0 × 0.6	45.34	23.63	2021	0.9805	3	15.17	26.25 ± 0.72	11.8 ± 1.0 (0.3)
PR03	Gneiss	Boulder	4.0 × 4.0 × 0.6	45.34	23.62	1970	0.996	3	14.33	13.61 ± 0.46	6.2 ± 0.5 (0.2)
PR04	Gneiss	Boulder	4.0 × 3.0 × 2.0	45.34	23.62	1976	0.9899	3	15.14	29.04 ± 0.77	13.4 ± 1.2 (0.3)
PR05	Gneiss	Boulder	2.0 × 1.5 × 1.0	45.34	23.62	2008	0.9929	3	10.60	19.64 ± 0.63	8.8 ± 0.2 (0.8)
PR06	Gneiss	Boulder	1.5 × 1.0 × 1.0	45.34	23.62	1908	0.9874	3	15.25	26.87 ± 0.60	13.1 ± 1.1 (0.3)
PR07	Gneiss	Boulder	2.5 × 1.5 × 0.5	45.34	23.62	1906	0.9652	3	10.61	26.12 ± 1.24	13.0 ± 1.2 (0.6)
PR08	Gneiss	Boulder	1.8 × 1.5 × 1.0	45.34	23.62	1905	0.9823	3	15.16	26.94 ± 0.64	13.2 ± 1.1 (0.3)
Gâlcescu cirque											
PR09	Gneiss	Boulder	3.5 × 2.0 × 3.0	45.35	23.61	1988	0.9897	3	9.70	28.37 ± 0.68	12.5 ± 1.1 (0.3)
PR10	Gneiss	Boulder	3.0 × 2.0 × 1.1	45.35	23.61	1990	0.9822	3	8.21	24.29 ± 0.69	11.2 ± 1.0 (0.3)
PR11	Gneiss	Boulder	2.0 × 1.6 × 2.0	45.35	23.61	1993	0.9844	3	8.91	26.84 ± 0.77	12.3 ± 1.1 (0.3)
PR12	Gneiss	Boulder	4.5 × 2.1 × 1.9	45.35	23.61	1941	0.9927	3	15.16	25.18 ± 0.68	11.9 ± 1.0 (0.3)
PR13	Gneiss	Boulder	7.0 × 4.0 × 4.0	45.35	23.61	1936	0.9916	3	15.21	32.16 ± 0.69	15.2 ± 1.3 (0.3)
PR14	Gneiss	Boulder	1.5 × 0.3 × 0.6	45.35	23.61	1932	0.9913	3	7.314	26.29 ± 0.83	12.5 ± 1.1 (0.4)
Zănoaga Mare cirque											
PR15	Quartz vein	Boulder	1.0 × 0.7 × 0.3	45.35	23.59	2055	0.9929	3	10.19	23.63 ± 0.70	10.2 ± 0.9 (0.3)
PR16	Gneiss	Boulder	2.0 × 1.0 × 1.0	45.35	23.59	2055	0.9929	3	10.56	23.94 ± 0.83	10.4 ± 0.9 (0.3)
PR17	Gneiss	Boulder	2.3 × 1.1 × 1.0	45.35	23.59	2047	0.99	3	10.81	28.49 ± 0.72	12.5 ± 1.1 (0.3)
PR18	Gneiss	Boulder	6.0 × 5.0 × 4.0	45.35	23.60	2043	0.9946	3	10.51	29.99 ± 1.00	13.1 ± 1.2 (0.4)
PR19	Gneiss	Boulder	4.4 × 2.4 × 3.1	45.35	23.60	2037	0.9957	3	3.06	30.97 ± 1.27	13.6 ± 1.3 (0.6)
PR20	Quartz vein	Boulder	2.5 × 1.6 × 1.0	45.35	23.60	2029	0.9948	3	10.07	32.20 ± 0.86	14.2 ± 1.3 (0.4)
PR21	Quartz vein	Boulder	3.0 × 2.7 × 1.5	45.35	23.60	2025	0.9864	3	10.65	31.06 ± 0.75	13.9 ± 1.2 (0.3)
Lotru valley											
PR22	Gneiss	Boulder	2.0 × 1.0 × 0.9	45.35	23.62	1769	0.9827	3	8.23	24.29 ± 0.73	13.1 ± 1.2 (0.4)
PR23	Gneiss	Boulder	3.0 × 2.5 × 2.0	45.35	23.62	1766	0.9819	3	11.18	24.16 ± 0.65	13.0 ± 1.1 (0.3)
PR24	Gneiss	Boulder	3.0 × 2.3 × 2.0	45.35	23.62	1769	0.9889	3	11.15	24.96 ± 0.66	13.4 ± 1.2 (0.4)

[a] Topographical shielding computed after Dunne et al. (1999).
[b] ^{10}Be concentrations. $^{10}Be/^{9}Be$ ratios measured against NIST SRM 4325 with nominal value of 2.79×10^{-11}. $^{10}Be/^{9}Be$ ratios were corrected by a full chemistry procedural blank that yielded 2.14–4.47% of the number of ^{10}Be atoms in the samples.
[c] Exposure ages calculated using Cronus-Earth $^{10}Be - ^{26}Al$ exposure age calculator v. 2.2 (http://hess.ess.washington.edu/). They assume zero erosion, scaling factors according to Stone (2000) and a spallation production rate of 4.49 ± 0.39 atom (g SiO_2) $^{-1}$ a^{-1} (Balco et al., 2008). Exposure ages are presented with the external uncertainties (internal uncertainties in brackets).

ridge of Coasta Păsări (Fig. 2). The amount and type of the debris indicate that the steep side walls are very active and provide lots of frost-shattered material to the talus cones below. On the south-east wall a large rock chute is the main contributor to the material accumulated below. Various glacial deposits are found at the bottom of this cirque and on the high step towards the Gâlcescu cirque, deposited as ice flew downwards. The Păsări Lake (>0.003 km^2) was formed between the glacial deposits and ice-moulded bedrock areas.

The Gâlcescu cirque is larger and located approximately 200 m below the Căldarea Dracului (Fig. 3b). The Gâlcescu cirque is surrounded by high steep walls in the southern and south-eastern part towards the Setea Mare peak and the Gâlcescu Ridge where talus cones are still actively forming. There are two glacially deepened basins: one basin occupied by Pencu Lake (0.002 km^2) and Vidal Lake (0.006 km^2) at the southern end of the cirque, and a wider basin with Gâlcescu Lake (0.03 km^2) towards the northern part (Figs. 2 and 3b). All lakes are limited by frontal glacial moraines at their northern end. On the eastern side of the lakes, lateral moraines were deposited along the cirque floor. They can be traced up along the eastern side of the cirque to the frontal moraines in front of the Gâlcescu Lake.

The Zănoaga Mare cirque is located at 2000–2050 m altitude between the Păsări Ridge to the south-east and Pietroasa Ridge to the north-west (Figs. 2 and 3d). There are widespread areas covered with frost-shattered material at the base of the steep cirque walls, especially in the southern and south-western part of this cirque, and partially covered in vegetation. The Zănoaga Mare cirque is the most representative for the glaciated environment in our study area. It is a wide and shallow cirque that was likely the major ice accumulation area feeding into the main Lotru valley glacier. Moreover, the large amounts of glacially polished areas on the lower south-eastern side walls and the elongated ice moulded bedrock are a good indicator of the direction and intensity of the ice flow (Figs. 2 and 3d).

The Zănoaga Mare cirque comprises one of the most representative erosional and depositional glaciated areas of the Southern Romanian Carpathians. Based on altitude and morphology, the Zănoaga Mare cirque can be divided into a western and an eastern compartment. The western compartment is smaller and located at higher altitude (~2100 m). Ice moulded bedrock in the form of roche moutonnées and whalebacks are typical for this part of the cirque (Fig. 3e). A lateral moraine deposited along the northern wall indicate the last ice flow towards the wider eastern compartment of the Zănoaga Mare cirque, located ~100 m lower (Fig. 2). It continues further down with another moraine that blocked Zănoaga Mare Lake (0.01 km^2). Towards the cirque headwall, a series of arched moraines stand as evidence for the last glacial activity in the Zănoaga Mare cirque. A well-developed protalus rampart is located in the upper part of the moraines (Fig. 3f). This arcuate ridge developed along the lower margin of a snow slope. Glacial deposits are widespread along the entire cirque floor and massive boulders of various dimensions have been abandoned on top of the deposits, especially in the middle part of the cirque. Two former ice flows are clearly visible in this lower part of the Zănoaga Mare cirque towards the Lotru Valley, separated by an elongated ice moulded bedrock feature (whaleback) (Fig. 2).

4.3. Lotru valley

The Lotru valley is formed at ~1670 m by the confluence of the Iezer and Gâlcescu streams (Fig. 2). Its high complexity in the

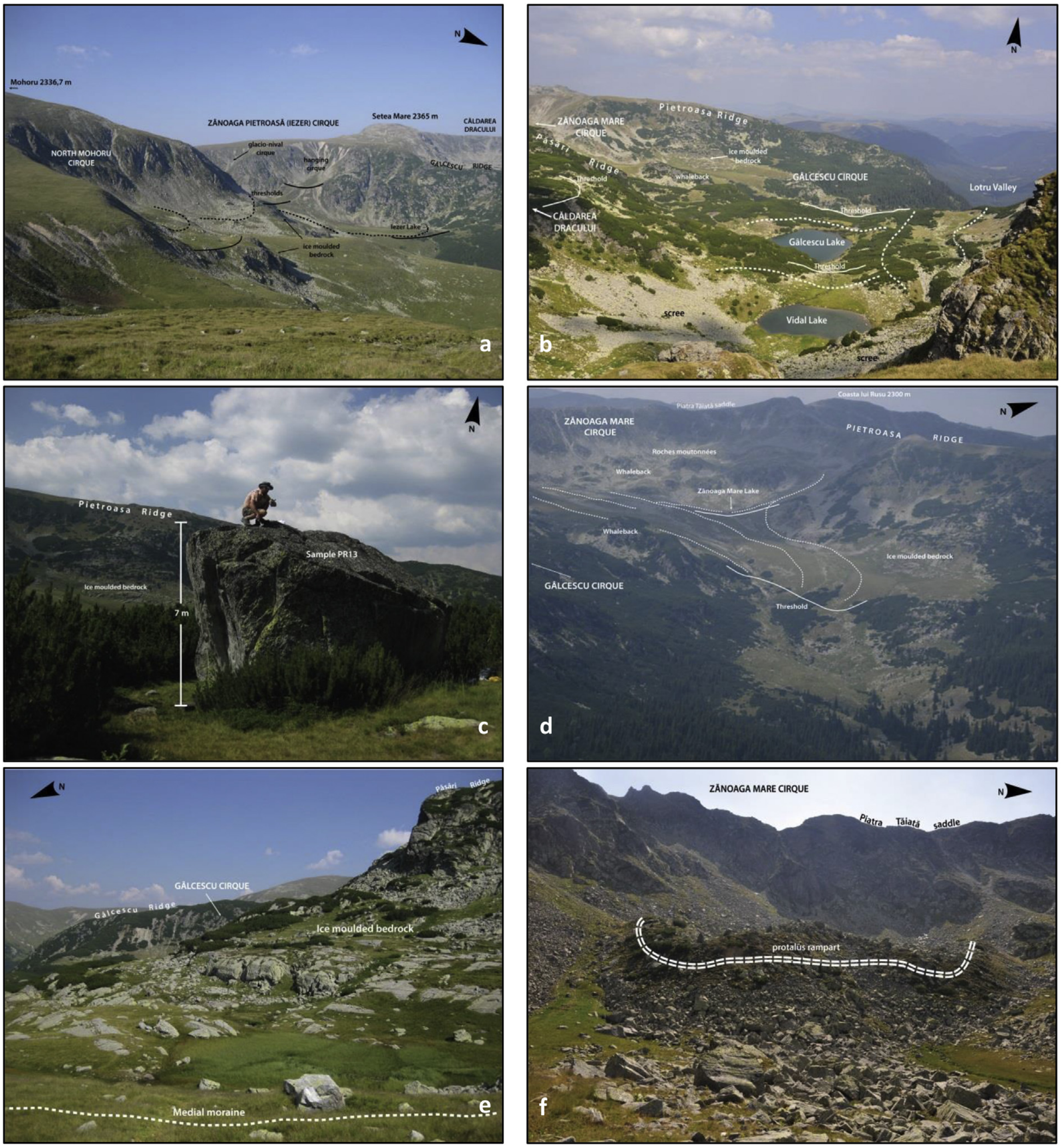

Fig. 3. Glaciated environment in the Parâng Mountains. Dashed lines indicate moraines, solid lines mark the glacial thresholds, double dashed lines delineate protalus ramparts. a) Iezer Valley with the Zănoaga Pietroasă and North Mohoru cirques; b) Moraines in the Gâlcescu cirque; c) Boulder sampled for surface exposure dating in the Gâlcescu cirque; d) Sediment-landform assemblages in the Zănoaga Mare cirque; e) Ice-moulded bedrock surfaces in the Zănoaga Mare cirque; f) Protalus rampart in the Zănoaga Mare cirque.

higher sector is due to the preglacial morphology, but also due to structural and lithological influence. The middle and lower sector of the Lotru valley is heavily forested, thus the glacial deposits are very difficult to identify. Unfortunately, ground truthing of most of the published glacial record at lower altitudes has been impossible after the building of a major storage reservoir at Vidra.

5. Results

The details of each sample collected in the Parâng Mountains and the 24 surface exposure ages are presented in Table 1. All surface exposure ages in Table 1 are presented with both uncertainties, however, in the discussion the internal uncertainty is used as all samples come from a small area with no significant difference in the ^{10}Be production rate.

Eight samples taken from three moraines in the cirques of the Iezer valley have a range of ^{10}Be ages from 13.4 ± 0.3 ka to 6.2 ± 0.2 ka. The down valley moraine in front of Iezer Lake (the Zănoaga Pietroasa cirque) yielded exposure ages of 13.1 ± 0.3 ka, 13.0 ± 0.6 ka and 13.2 ± 0.3 ka. The probability density plot suggests a tight cluster around a weighted mean of 13.2 ± 0.3 ka (Fig. 4). Further up, the moraine located along the Iezer eastern valley wall yielded exposure ages of 6.2 ± 0.2 ka, 13.4 ± 0.3 ka and 8.8 ± 0.8 ka. The scatter of exposure ages confirms the mixed origin of this landform. The oldest age of 13.4 ± 0.3 ka was produced from a boulder located further away from the walls in a stabile area of the landform. We tentatively assigned a deglaciation age consistent with the rest of the ages produced from samples located in front of the Iezer Lake (see above). The two younger ages of 6.2 ± 0.2 ka and 8.8 ± 0.8 ka are most likely caused by rockfall from the steep side wall. Boulders on another moraine in the North Mohoru cirque were deposited at 11.2 ± 0.5 ka and 11.8 ± 0.3 ka with a weighted mean of 11.8 ± 0.2 ka (Figs. 4 and 5).

Six samples were collected from around the lakes in the Gâlcescu cirque with ages suggesting deposition at 12.5 ± 0.3 ka, 11.2 ± 0.3 ka, 12.3 ± 0.3 ka, 11.9 ± 0.3 ka, 15.2 ± 0.3 ka and 12.5 ± 0.4 ka (Fig. 5). Given the size of the boulder (7 m a-axis), it is very likely that the boulder deposited at 15.2 ± 0.3 ka has been exposed to cosmic radiation before deposition, thus explaining the high ^{10}Be concentration and exposure age in comparison with the rest of the ages. Pre-exposure on a rock cliff and transport on top of ice surface are likely the causes for the inheritance of nuclides in this boulder, here considered an outlier. According to the probability density plot the other five ages are tightly clustered around a weighted mean age of 12.6 ± 0.3 ka (Figs. 4 and 5).

Above 2000 m altitude, seven boulders spread throughout the Zănoaga Mare cirque were exposure dated to 10.2 ± 0.3 ka, 10.4 ± 0.3, 12.5 ± 0.3 ka, 13.1 ± 0.4 ka, 13.6 ± 0.6 ka, 14.2 ± 0.4 ka and 13.9 ± 0.3 ka (Fig. 5). The last five samples produced from the moraines in this cirque come from a widespread moraine complex and it is unlikely to have been covered. These samples come from boulders well-embedded into the moraine surfaces and sticking out well above the landform (>1 m). The younger two ages of 10.2 ± 0.3 ka and 10.4 ± 0.3 could have been covered as they were collected from a subdued moraine, thus indicating the time of stabilization of the moraine rather than its deposition.

Three samples from boulders in the Lotru valley yielded exposure ages of 13.1 ± 0.4 ka, 13.0 ± 0.3 ka and 13.4 ± 0.4 ka (Fig. 5). The probability density plot suggests a tight cluster around a weighted mean of 13.2 ± 0.3 ka (Figs. 4 and 5). All boulders were well embedded into the surface and unlikely to have suffered any modifications since their deposition.

6. Discussion

6.1. Chronology of deglaciation in the Parâng Mountains

During the last glaciation (Würm, 125–11 ka), the Carpathian cirques were excavated and deepened by glaciers. They were subsequently filled with glacial sediments which stand as evidence of a reshaped and very diverse landscape. Previous studies in the Parâng Mountains identified repeated glaciers advances and recessions during the last Quaternary glaciations (e.g. Iancu, 1972; Vuia, 2002). In this part we present our interpretation of field evidence and the deglaciation chronology obtained with the surface exposure dating of moraine boulders.

We found no evidence of glacial deposits in the lower Lotru valley during our field season (below 1700 m altitude). However, this valley has been extensively modified through incision of the Lotru river, especially in its middle sector (especially around 1400 m altitude), and any evidence was likely removed during the Holocene period. No suitable sites for cosmogenic nuclide surface exposure dating were found here.

During the Late Glacial, the downwasting of ice in the upper valleys and cirques of the Parâng Mountains (Figs. 2 and 5) occurred after 14 ka when the climate became warmer (Björck et al., 1998). At 13.2 ± 0.3 ka, boulders were abandoned at ~1700 m altitude in the upper Lotru valley (PR22, 23, 24), soon after separation from the ice flowing from the Gâlcescu and Zănoaga Mare cirques. These boulders are part of a deposit likely formed by the glacier flowing down from the North Mohoru cirque (Figs. 2 and 5). At roughly the same time, ice in the Iezer valley had already withdrawn towards the higher ground. In the Zănoaga Pietroasa cirque, this southward retreat of glacier caused the deposition of the moraine that blocked

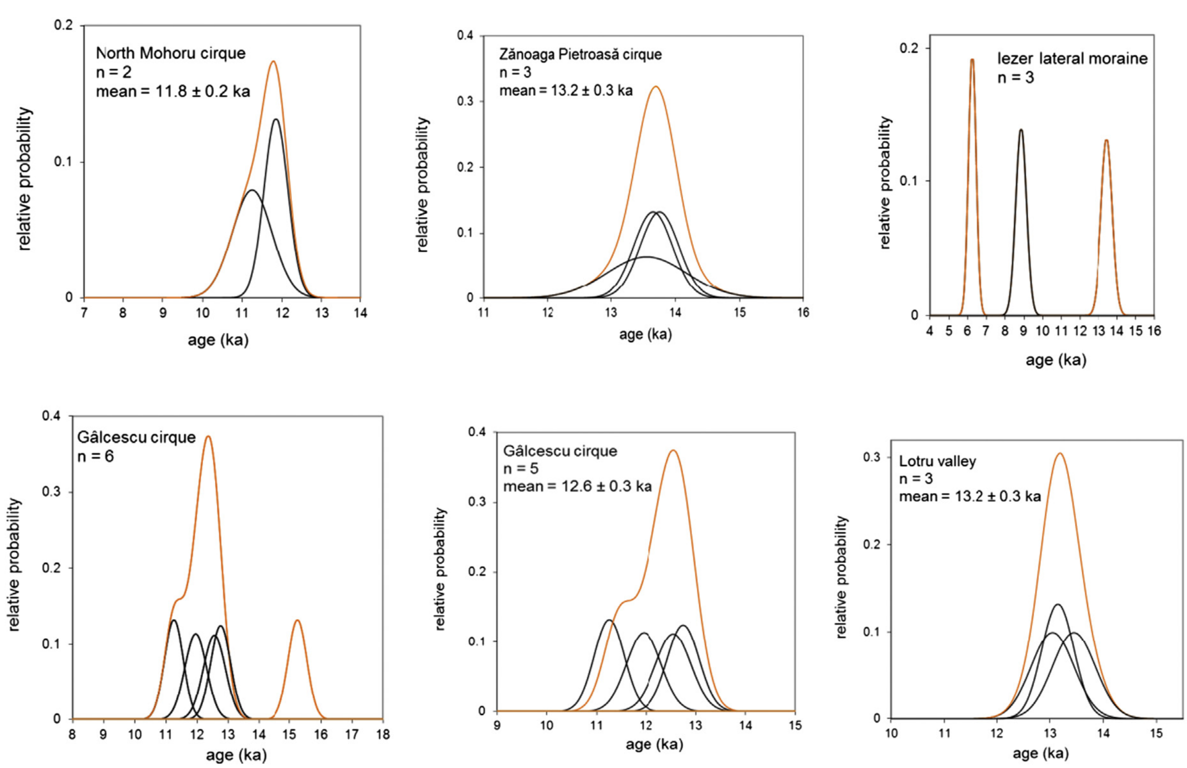

Fig. 4. Probability density plots for the exposure ages obtained from glacial landforms. The ages are given with their analytical uncertainty.

the Iezer Lake at 13.2 ± 0.3 ka (PR06, 07, 08). Meanwhile, the glacier was not only retreating upwards but was also gradually narrowing towards the middle of the valley, as suggested by the deposition of the lateral moraine at 13.4 ± 0.3 ka (PR04). This lateral moraine is covered with various boulders, mostly embedded in its surface especially towards its northern part (Fig. 2). The two younger ages of 6.2 ± 0.2 ka and 8.8 ± 0.8 ka were produced from samples collected further south nearer to the side wall. It is well documented that rock slope failures often occurred during the Holocene as a result of landscape readjustment after ice retreat (Ballantyne and Stone, 2013; Ballantyne et al., 2014).

A faster retreat of ice occurred from the Lotru valley towards the Zănoaga Mare cirque. Boulders on the lateral moraine below the Zănoaga Mare Lake were abandoned at 14.2 ± 0.4 ka and 13.9 ± 0.3 ka (PR20, 21). Ice in this wide and shallow cirque would have received more solar radiation than the others cirques, speeding up the glacier retreat after separation from the Gâlcescu glacier.

The subsequent thinning and narrowing of ice towards the middle of the cirque is indicated by the erosional paths of two former ice flow that are still visible today around the large whaleback located on the glacial threshold (Figs. 2 and 3b,d). One of the ice flows followed the main Zănoaga stream (on the north side) while the secondary flow was probably joined to the Gâlcescu glacier. During this time, a large amount of boulders were abandoned on top of the elongated deposits located in the middle of the cirque floor at 13.6 ± 0.6 ka, 13.1 ± 0.4 ka and further up at 12.5 ± 0.3 ka (PR19, 18, 17) as ice was slowly retreating towards the cirque's headwall (Figs. 2 and 3e).

Although the Gâlcescu cirque had a smaller accumulation area, snow drift from the upper surfaces would have contributed to a thicker glacier (Gellatly et al., 1989; Payne and Sugden, 1990; Mitchell, 1996). Also sheltered by steep high walls, the Gâlcescu ice retreated slower than the glacier in the Zănoaga Mare cirque located at similar altitudes. This is supported by the weighted mean ^{10}Be exposure age of 12.6 ± 0.3 ka (PR09, 10, 11, 12, 14) produced

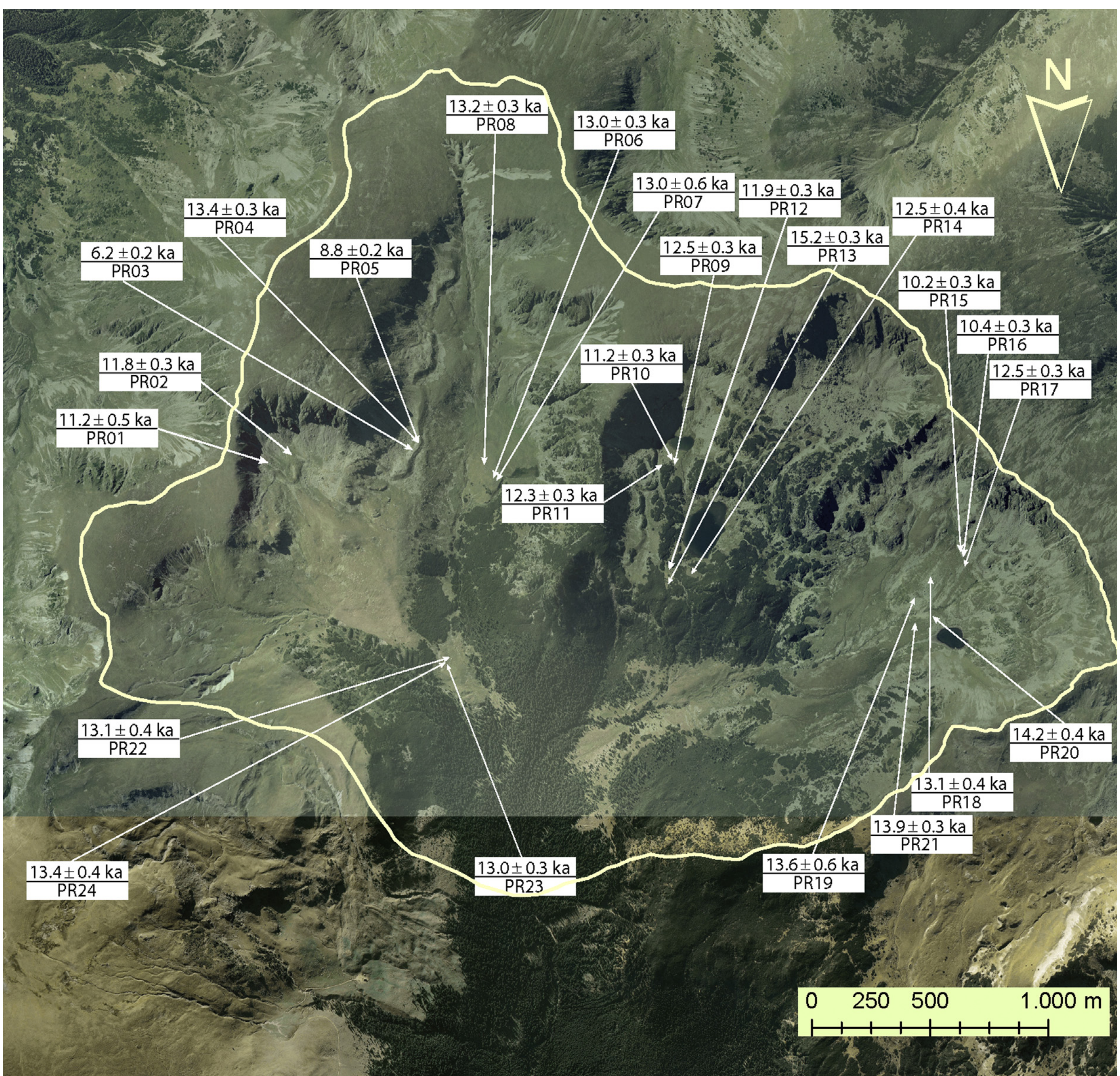

Fig. 5. Surface exposure ages (with analytical uncertainties) and their positions in the study area.

from the boulders deposited in this area, indicating that the cirque basin was free of glacial ice by ca. 11.0 ka (Figs. 2 and 5). The glacier located in the Căldarea Dracului cirque was probably feeding the glacier that occupied the Gâlcescu cirque. Though no glacial landforms were exposure dated in the Căldarea Dracului cirque, its higher location (~200 m above) would have caused the maintenance of ice much longer than in the other cirques. Moreover, the final retreat of the glaciers from the upper Late Glacial moraine in the North Mohoru cirque occurred at 11.8 ± 0.2 ka (PR01, 02), marking the beginning of the Holocene warming period (Figs. 2 and 3a).

Following more retreat, the glacier in the Zănoaga Mare cirque deposited the recessional moraines near the southern headwall (Figs. 2 and 3f). Here, the glaciers finally reached smaller dimensions and melted completely at the beginning of the Holocene as suggested by the ages from the boulders in the innermost moraines at 10.2 ± 0.3 ka and 10.4 ± 0.3 ka (PR15, 16). The higher western compartment of the Zănoaga Mare cirque likely preserved ice for longer times, similar to the Căldarea Dracului cirque. Based on the ^{10}Be ages in this study, the final downwasting of the Late Würmian glaciers in this part of the Parâng Mountains took no more than a few thousand years (between 14 and 10 ka), especially with the warmer Holocene climate as an important mechanism for enhanced ablation.

6.2. ELA reconstruction

Based on the glacial geomorphology and the chronology constrained using surface exposure ^{10}Be ages (Fig. 2 and Table 1), we estimated the equilibrium line altitudes (ELAs) for the cirque glaciers in the central area of the Parâng Mountains. Generally, the ELAs vary greatly during a glacial stage, depending on the climatic conditions. In the case of a more stable climate, the ELA remains constant when moraine deposition occur (Hawkins, 1985; Matthews, 2013). We calculated the ELA using the accumulation area ratio (AAR) with a value of 0.65 ± 0.5 as the percentage of a glacier's accumulation area above the ELA (Porter, 1975, 2001).

Given the small size of the study area and the similar altitudes of the cirque floors, the ELA did not vary greatly between the neighbouring valleys and cirques. The ELA was located at 1990 m in both the Iezer valley and the Gâlcescu cirque at ~13 ka. In the Zănoaga Mare glacier, wider and shallower than the other two cirques, the ELA was established at 2027 m during the Late Glacial. This agrees well with the reconstructed ELAs at ~2030 m in the Retezat Mountains, west of the Parâng Mountains (Reuther et al., 2007) while Kuhlemann et al. (2013a) suggest ELA values between 1950 and 2130 m in the central part of the Făgăraş Mountains for the Late Glacial period. Further north in the Rodna Mountains, Gheorghiu (2012) estimated the ELA at ~1800 m for the Late Glacial period. There, a lower ELA is generally explained through a more extensive ice advance due to its more northerly position and closer to the cold air from the Baltic and the Siberian regions (Gheorghiu, 2012).

6.3. Comparison with other Romanian records

Our results suggest that the alpine areas of the Southern Romanian Carpathians experienced deglaciation at the same time (14–10 ka) as the other areas located west (Retezat Mountains – Reuther et al., 2007; Făgăraş Mountains – Kuhlemann et al., 2013a) and further north (Rodna Mountains – Gheorghiu, 2012) during the Late Glacial period. Although evidence of glaciation exists in the middle and lower Lotru valley, no boulders or bedrock areas were found suitable for surface exposure dating of the maximum ice extent in the Parâng Mountains. Moreover, based on the current morphological characters of the study area, it was not possible to establish whether a climatically controlled readvance occurred during the Late Glacial or the Younger Dryas, or whether the deglaciation occurred as a continuous event since the maximum ice extent.

The retreat of ice towards the higher valleys and cirques in the Parâng Mountains during the Late Glacial period is indicated by the exposure ages obtained from moraine boulders. Glacial retreat in the Iezer valley, and the Gâlcescu and Zănoaga Mare cirques by ~14 ka (Figs. 2 and 6) may coincide with a moderate warming of the climate corresponding to the warming recorded in the Greenland ice cores (GI-1e, 14.7 ka; Björck et al., 1998) (Bølling interstadial). A warming period was recorded in speleothems in the Bihor Mountains at 14.8 ka (Tămaş et al., 2005). Braun et al. (2012) established a subsequent well-defined period of warmer conditions between 14.2 and 13.9 ka in the Retezat Mountains. At this time, glaciers in the Retezat Mountains (west of the study area) had already retreated in the sheltered cirques (12–10 ka, recalculated here as equivalent to ~14–12 ka; Reuther et al., 2007). Similarly, at higher latitudes in the Rodna Mountains (255 km north), withdrawal of ice in the northern cirques occurred by 14–13 ka (Gheorghiu, 2012). This is in good agreement with the warm and dry climate found at lower altitudes in north-western Romania (Wohlfarth et al., 2001). Increasingly warmer conditions are also reported in the Gutâi Mountains between 13.8 and 12.9 ka with a clear altitudinal shift of vegetation towards higher grounds (Björkman et al., 2002).

Only weak cooling was observed in chironomid reconstruction during the Younger Dryas episode (Tóth et al., 2012) in sediments cored from the Brazi Lake, Retezat Mountains (west of Parâng). However, ^{10}Be exposure ages from the three cirques in the central Parâng Mountains indicate the presence of ice during the cold period of the Younger Dryas after 13 ka (Fig. 6). Our ages are consistent with a deglaciation age of 12.8 ± 2.0 ka in the Făgăraş Mountains (Kuhlemann et al., 2013a). The cold Younger Dryas stage at higher altitudes was recorded in the Bihor Mountains (western Romania) between 12.6–11.4 ka (Tămaş et al., 2005).

The change from cold to warm conditions at 11.7 ka corresponds to the Late Glacial/Holocene transition (Björck et al., 1998). Timing of the glacial retreat in the Iezer North Mohoru cirque (11.8 ± 1.1 ka; Fig. 6) is consistent with the rapid growth of speleothems after ~11.5 ka in the Poleva cave (300 km south-west) (Constantin et al., 2007). Final deglaciation ages in the study area (10.2 ± 1.0 ka and 10.4 ± 0.9 ka; Fig. 6) correlate well with the Holocene vegetation expansion at the beginning of the warm period based on pollen analysis in the Avrig area (75 km north-east) (Tanţău et al., 2006).

6.4. Glacial retreat compared to other records in Europe

Similar climatic changes occurred in the mountains of Europe during the Late Glacial period. Retreat of ice from the maximum position occurred between 16.7 ± 2.3 ka and 10.8 ± 1.4 ka in the Šara range of the Balkan Peninsula (Kuhleman et al., 2009;) and earlier in the Rila Mountains (18–16 ka; Kuhleman et al., 2013b). The deposition of boulders in the Parâng Mountains between 14.2 and 10.2 ka is supported by other published data. Based on $\delta^{18}O$ on bulk carbonates from lake sediments in Slovenia an increase of temperatures was recorded at 14.8 ka (Andrič et al., 2008). The Late Glacial moraines in central Turkey have been exposure dated to 14.6 ± 1.2 ka (Sarikaya et al., 2009). The high altitude areas of the Tatra Mountains also experienced significant loss of ice volume before 13 ka during the warming of the Bølling/Allerød interstadial (14.7–12.7 ka) with a final deglaciation at 9.4 ± 2.1 ka (Makos et al., 2013). Additionally, surface exposure dating of a moraine in the Ukrainian Carpathians indicates a Younger Dryas deposition with a mean age of 12.1 ± 0.3 ka (Rinterknecht et al., 2012). The last major

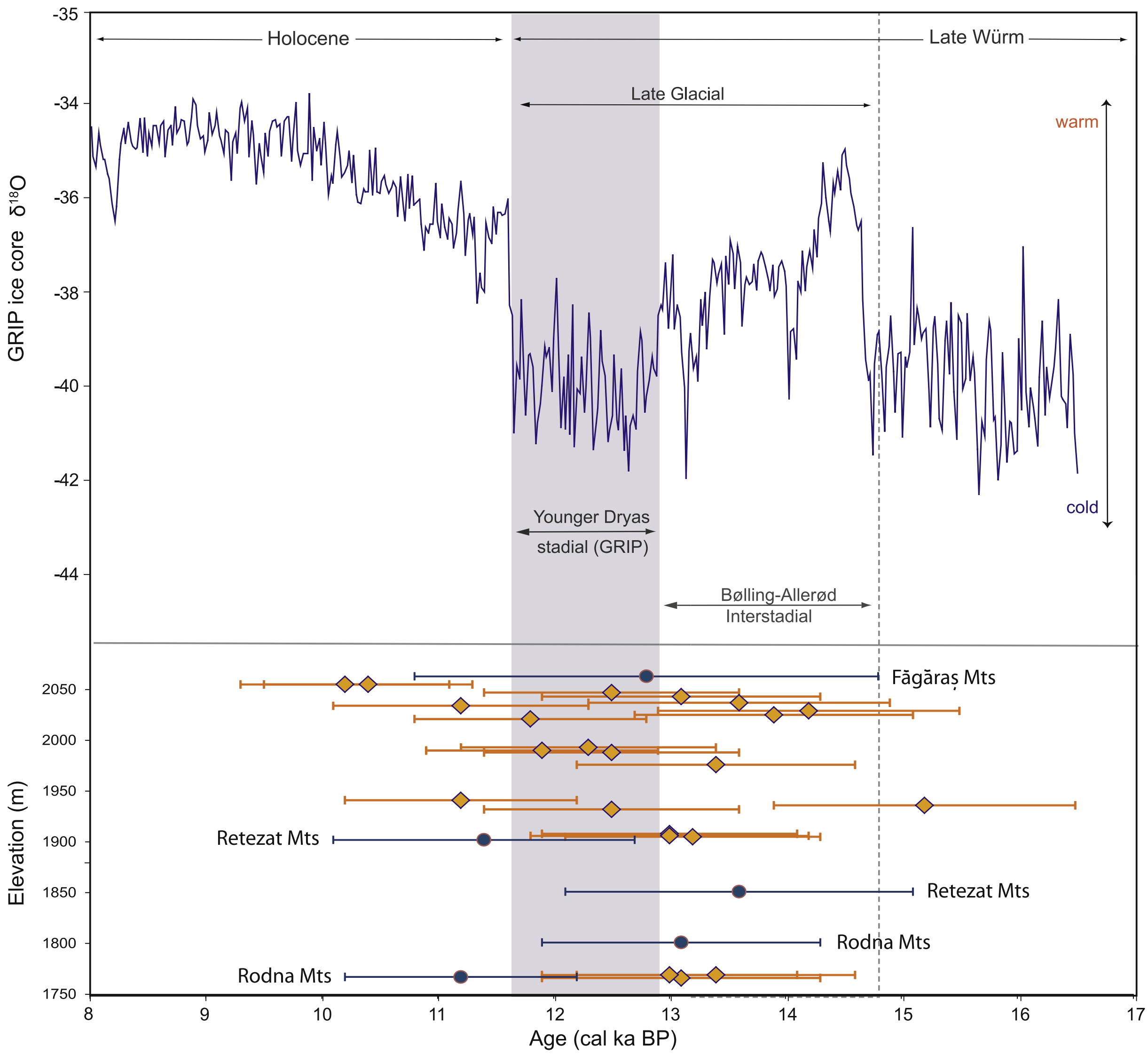

Fig. 6. ^{10}Be surface exposure ages (with systematical uncertainties) plotted against the GRIP ice core (Rasmussen et al., 2006). Several ^{10}Be exposure dates from Romanian Carpathians are included for comparison with our data.

cooling phase before the warm Holocene was also recorded in the sediments of the Przedni Staw Lake in the Northern Carpathians (Lindner et al., 2003). The Younger Dryas cirque glaciers at high elevations in the Eastern Alps retreated at ca. 11.5 ka (van Husen, 1997; Ivy-Ochs et al., 2006, 2008) and other evidence from southern Turkey confirms melting of ice at ca 11.5 ka (Zahno et al., 2010). Similar deglaciation occurred synchronously in the Albanian Alps and northern Greece (Hughes and Braithwaite, 2008; Hughes and Woodward 2008; Milivojevič et al., 2008).

7. Conclusions

Comparison of our reported ages to other chronologies within the region supports an interpretation that our data are accurate within the context of the study location, the outlined glacial stratigraphy, and within the context of latest Pleistocene climate change. The results presented in this study are consistent with cosmogenic nuclide exposure ages previously reported from adjacent areas. The ^{10}Be exposure ages in this study range between 14.2 and 10.2 ka and indicate the deglaciation time of late Würmian glacial complexes within the central Parâng Mountains. The ELA during the Late Glacial was situated at 1990 m altitude, consistent with other glaciated areas of the Carpathians.

The Late Glacial retreats in the study area also occurred simultaneously with periods of cold and wet climatic conditions in the North Atlantic region. Being located further away from the North Atlantic, a different climatic signal is more likely expected. However, the ^{10}Be exposure ages confirm that ice was still maintained until the beginning of the Holocene in the higher parts of the Carpathians. Deglaciation of the upper valleys and cirques, and thus the entire glacial system in the Parâng Mountains was complete by approximately 10 ka. Additional chronologies for deglaciation within the region are necessary to resolve this issue and also to establish and constrain the maximum extent of ice.

References

Administrația Națională de Meteorologie –ANM, 2008. Clima Romaniei. In: Bucuresti Bercia, I., Codarcea, Al, Marinescu, Fl (Eds.), 1967. Geological Map of Romania. Editura Academiei Române, 1:200.000, Sheet 33, Târgu Jiu.

Akçar, N., Yavuz, V., Ivy-Ochs, S., Kubik, P.W., Vardar, M., Schlüchter, C., 2007. Paleoglacial records from Kavron Valley, NE Turkey: field and cosmogenic exposure dating evidence. Quaternary International 164–165, 170–183.

Andrič, M., Kroflič, B., Toman, M.J., Ogrinc, N., Dolenec, T., Dobnikar, M., Čermelj, B., 2008. Late quaternary vegetation and hydrological change at Ljubljansko barje (Slovenia). Palaeogeography Palaeoclimatology Palaeoecology 270, 150–165.

Balco, B., Stone, J., Lifton, N., Dunai, T., 2008. A simple, internally consistent, and easily accessible means of calculating surface exposure ages and erosion rates from Be-10 and Al-26 measurements. Quaternary Geochronology 3, 174–195.

Ballantyne, C.K., Stone, J.O., 2013. Timing and periodicity of paraglacial rock-slope failures in the Scottish Highlands. Geomorphology 186, 150–161.

Ballantyne, C.K., Wilson, P., Gheorghiu, D., Rhodes, A., 2014. Enhanced rock-slope failure following ice-sheet deglaciation: timing and causes. Earth Surface Processes and Landforms 39, 900–913.

Bercia, I., Marinescu, F.l, Mutihac, V., Pavelescu, M., Stancu, I., 1967. The Geological Map of the Parâng Mountains, Sheets 26 Orăştie and 33 Târgu Jiu. Geological and Geophysics Institute of Romania.

Björck, S., Walker, M.J.C., Cwynar, L.C., Johnsen, S., Knudsen, K.L., Lowe, J.J., Wohlfarth, B., INTIMATE Members, 1998. An event stratigraphy for the last termination in the North Atlantic region based on the Greenland ice-core record: a proposal by the INTIMATE group. Journal of Quaternary Science 13, 283–292.

Björkman, L., Feurdean, A., Cinthio, K., Wohlfarth, B., Possnert, G., 2002. Lateglacial and early Holocene vegetation development in the Gutâiului Mountains, NW Romania. Quaternary Science Reviews 21, 1039–1059.

Braun, M., Hubay, K., Magyari, E., Papp, I., Bálint, M., 2012. Using linear discriminant analysis (LDA) of bulk sediment geochemistry data to reconstruct lateglacial climate change in the South Carpathians. Quaternary International 293, 114–122.

Child, D., Elliott, G., Mifsud, C., Smith, A.M., Fink, D., 2000. Sample processing for earth science studies at ANTARES. Nuclear Instruments and Methods in Physics Research Section B: Beam Interactions with Materials and Atoms 172, 856–860.

Constantin, S., Bojar, A.M., Lauritzen, S.E., Lundberg, J., 2007. Holocene and Late Pleistocene climate in the sub-Mediterranean continental environment: a speleothem record from Poleva Cave (Southern Carpathians, Romania). Palaeogeography, Palaeoclimatology, Palaeoecology 243, 322–338.

Drăguţ, L., 2004. Munţii Şureanu. Studiu geomorfologic. teză de doctorat. Facultatea de Geografie, Universitatea Babeş-Bolyai, Cluj-Napoca.

Dunne, J., Elmore, D., Muzikar, P., 1999. Scaling factors for the rates of production of cosmogenic nuclides for geometric shielding and attenuation at depth on sloped surfaces. Geomorphology 27, 3–11.

Florineth, D., Schlüchter, C., 1998. Reconstructinglast Glacial Maximum (lgm) ice surface geometry and flowlines in the central swiss alps. Eclogae Geologicae Helvetiae 91, 391–407.

Gellatly, A.F., Whalley, W.B., Gordon, J.E., Hansom, J.D., Twigg, D.S., 1989. Recent glacial history and climatic change, Bergsfjord, Troms-Finnmark, Norway. Norsk geographie Tidsskrift 43, 19–30.

Gheorghiu, D., 2012. Testing Climate Synchronicity since the Last Glacial Maximum between Scotland and Romania. PhD thesis. University of Glasgow, p. 200.

Gillespie, A., Molnar, P., 1995. Asynchronous maximum advances of mountain and continental glaciers. Reviews of Geophysics 33, 311–364.

Grozescu, M.H., 1920. Morphologie de la vallée du Lotru, Comptes-Rendus des Séances, T, VIII. Institut Geologique de Roumanie, Bucharest, pp. 178–181.

Hawkins, F.F., 1985. Equilibrium-line altitudes and palaeoenvironment in the merchants bay area, Baffin Island, N.W.T., Canada. Journal of Glaciology 31.

Hughes, P.D., Woodward, J.C., 2008. Timing of glaciation in the Mediterranean mountains during the last cold stage. Journal of Quaternary Science 23, 575–588.

Hughes, P.D., Woodward, J.C., Gibbard, P.L., 2006. Late Pleistocene glaciers and climate in the Mediterranean region. Global and Planetary Change 46, 83–98.

Hughes, P.D., Woodward, J.C., van Calsteren, P.C., Thomas, L.E., Adamson, K.R., 2010. Pleistocene ice caps on the coastal mountains of the Adriatic Sea. Quaternary Science Reviews 29 (27–28), 3690–3708.

Iancu, S., 1958. Câteva aspecte litologice şi structurale în morfologia glaciară a Masivului Parâng, Natura. Seria Geografie-Geologie X (1), 59–65.

Iancu, S., 1972. Treptele din profilul longitudinal şi cel transversal al văilor glaciare din Munţii Parâng, Lucr. Simp. de geografie fizică a Carpaţilor, (Bucureşti-Septembrie 1970). Inst. de Geografie al Academiei, pp. 111–116.

Iancu, S., 1963. Consideraţii asupra formării circurilor glaciare în trepte. Probleme de geografie X, 231–238.

Kuhlemann, J., Milivojević, M., Krumrei, I., Kubik, P.W., 2009. Last glaciation of the Šara range (Balkan peninsula): increasing dryness from the LGM to the Holocene. Austrian Journal of Earth Sciences 102, 146–158.

Kohl, C.P., Nishiizumi, K., 1992. Chemical isolation of quartz for measurement of in situ-produced cosmogenic nuclides. Geochimica et Cosmochimica Acta 56, 3586–3587.

Kuhlemann, J., Dobre, F., Urdea, P., Krumrei, I., Gachev, E., Kubik, P., Rahn, M., 2013a. Last glacial maximum glaciation of the Central south carpathian range (Romania). Austrian Journal of Earth Sciences 106, 83–95.

Kuhlemann, J., Gachev, E., Gikov, A., Nedkov, S., Krumrei, I., Kubik, P., 2013b. Glaciation in the Rila Mountains (Bulgaria) during the last glacial maximum. Quaternary International 293, 51–62.

Lehmann, P.W., 1881. Beobachtungen über Tektonik und Gletscherspuren im Fogaraschen Gebirge. Zeitschrift Geologie Gesellschaft Berlin 33, 109–117.

Lehmann, F.W.P., 1903. Schneeverhältnisse und Gletscherspuren in den Transsylvanischen Alpen. Jahresbericht der geographischen Gesellschaft Greifswald 9, 1–26 (in German).

Lindner, L., Dzierźek, J., Marciniak, B., Nitychoruk, J., 2003. Outline of Quaternary glaciations in the Tatra Mts.: their development, age and limits. Geological Quarterly 47 (3), 269–280. Warszawa.

Lişteveanu, P.C., 1942. Valea glaciară a Lotrului, Rev. geografică română V (I-II), 38–42.

Makos, M., Nitychoruk, J., Zreda, M., 2013. Deglaciation chronology and paleoclimate of the Pięciu Stawów Polskich/Roztoki Valley, high Tatra Mountains, Western Carpathians, since the Last Glacial Maximum, inferred from 36Cl exposure dating and glacier–climate modeling. Quaternary International 293, 63–78.

Marjanac, L., Marjanac, T., 2004. Glacial history of the Croatian Adriatic and Coastal Dinarides. In: Ehlers, J., Gibbard, P.L. (Eds.), Quaternary Glaciations – Extent and Chronology. Part I: Europe. Elsevier, Amsterdam, pp. 19–26.

de Martonne, Emm., 1907. Recherches sur l'évolution morphologique des Alpes de Transylvanie (Karpates méridionales). Paris, 286 p, vezi şi, Lucrări geografice despre România. In: Rev. de géogr. annuelle, t. I (1906–1907), vol. I, p. 271. Edit. Academiei, Bucureşti, 1981.

Matthews, J., 2013. Encyclopedia of Environmental Change. SAGE Publications Ltd., p. 1496

Milivojević, M., Menković, L., Ćalić, J., 2008. Pleistocene glacial relief of the central part of Mt. Prokletije (Albanian Alps). Quaternary International 190, 112–122.

Mitchell, W.A., 1996. Significance of snowblowing in the generation of Loch Lomond Stadial (Younger Dryas) glaciers in the Western Pennines, northern England. Journal of Quaternary Science 11, 233–248.

Morariu, T., 1981. Le relief glaciaire des Carpates Orientales de Roumanie. Rech. Géographique a Strasbourg 16–17, 67–69.

Pawłowski, St, 1936. Les Karpates á l'époque glaciaire. In: C. R. Congr. Intern. Géogr., Varsovie,Travaux de la section II, II, pp. 89–141.

Payne, A., Sugden, D.E., 1990. Topography and ice sheet growth. Earth Surface Processes and Landforms 15, 625–639.

Peltier, W.R., Fairbanks, R.G., 2006. Global glacial ice volume and Last Glacial Maximum duration from an extended Barbados sea level record. Quaternary Science Reviews 25, 3322–3337.

Peltier, W.R., Vettoretti, G., Stastna, M., 2006. Altantic meridional overturning and climate response to Arctic Ocean freshening. Geophysical Research Letters 33. http://dx.doi.org/10.1029/2005GL025251.

Posea, G., 1974. Relieful României. Editura Ştiinţifica, p. 483. Bucureşti.

Posea, G., 2003. Geografia fizică a României (Date generale, Pozi021Bia geografică, Relief). Editura Fundaţiei România de Mâine, p. 262. Partea I.

Posea, G., 2004. Geografia fizică a României (Clima, Apele, Biogeografia, Solurile, Hazarde natural). Editura Fundaţiei România de Mâine, p. 234. Partea a II-a.

Porter, S.C., 1975. Equilibrium line altitudes of late Quaternary glaciers in the Southern Alps. New Zealand. Quaternary Research 5, 27–47.

Porter, S., 2001. Snowline depression in the tropicsduring the Last Glaciation. Quaternary Science Reviews.

Rasmussen, S.O., Andersen, K.K., Svensson, A.M., Steffensen, J.P., Vinther, B.M., Clausen, H.B., Siggaard-Andersen, M.-L., Johnsen, S.J., Larsen, L.B., Dahl-Jensen, D., Bigler, M., Rothlisberger, R., Fischer, H., Goto-Azuma, K., Hansson, M.E., Ruth, U., 2006. A new Greenland ice core chronology for the last glacial termination. Journal of Geophysical Research 111. D06102.

Reuther, A.U., Urdea, P., Geiger, C., Ivy-Ochs, S., Niller, H.P., Kubik, P.W., Heine, K., 2007. Late Pleistocene glacial chronology of the Pietrele Valley, Retezat Mountains, Southern Carpathians constrained by 10Be exposure ages and pedological investigations. Quaternary International 164–165, 151–169.

Rinterknecht, V., Matoshko, A., Gorokhovich, Y., Fabel, D., Xu, S., 2012. Expression of the Younger Dryas cold event in the Carpathian Mountains, Ukraine? Quaternary Science Reviews 39, 106–114.

Sârcu, I., Sficlea, V., 1956. Cîteva observaţii geomorfologice în Munţii Parîngului şi ai Şureanului. Annals, University Al. I. Cuza" Iaşi, (Serie nouă), secţ. II (Şt. nat.-geografie) II (2), 384–402.

Sârcu, 1963. Le probleme de laglaciation quaternaire dans les montagnes du Maramureş. Anal. şt. Univ., Al. I. Cuza" Iaşi (Serie nouă), secţ. II(Şt. nat.) b. In: Geologie-Geografie, t. IX, pp. 125–134. I.

Sarıkaya, M.A., Zreda, M., Çiner, A., 2009. Glaciations and paleoclimatic variations on Mount Erciyes, central Turkey, since the Last Glacial Maximum, inferred from ^{36}Cl cosmogenic dating and glacier modeling. Quaternary Science Reviews 28, 2326–2341.

Savu, H., Dessila-Codarcea, M., Pavelescu, M., 1968. Geological Map of Romania, 1: 200.000, Sheet 26, Orăştie.

Stone, J.O. 2000. Air pressure and cosmogenic isotope production. Journal of geophysical research 105, No. B10, pages 23,753-23,759.

Tanţău, I., Reille, M., Beaulieu, J.L., Fărcaş, S., 2006. Late Glacial and Holocene vegetation history in the southern part of Transylvania (Romania) pollen analysis of two sequences from Avrig. Journal of Quaternary Science 21, 49–61.

Tóth, M., Magyari, E.K., Brooks, D.S.J., Braun, M., Buczkó, K., Bálint, M., Heiri, O., 2012. A chironomid-based reconstruction of lateglacial summer temperatures in the Southern Carpathians (Romania). Quaternary Research 77, 122–131.

Tămaş, T., Onac, B.P., Bojar, A.V., 2005. Lateglacial-Middle Holocene stable isotope records in two coeval stalagmites from the Bihor Mountains, NW Romania. Geological Quarterly 49, 185–194.

Urdea, P., Drăguţ, L., 2000. Noi date asupra reliefului glaciar şi periglaciar din Munţii Şureanu. Studii şi cercetări de geografie 47, 40–53.

Urdea, P., Reuther, A.U., 2009. Some new data concerning the Quaternary glaciation in the Romanian Carpathians. Geographica Pannonica 13 (2), 41–52.

Urdea, P., Onaca, A., Ardelean, F., Ardelean, M., 2011. New evidence on the Quaternary glaciation in the Romanian Carpathians. In: Ehlers, J., Gibbard, P.L., Hughes, P.D. (Eds.), Quaternary Glaciations – Extent and Chronology. Elsevier, Amsterdam.

van Husen, D., 1997. LGM and Late Glacial fluctuations in the Eastern Alps. Quaternary International 38 (39), 109–118.

Vuia, F., 2002. Diferenţieri morfometrice şi morfografice ale circurilor glaciare din bazinul Jieţului (Munţii Parâng). In: Analele Universităţii de Vest din Timişoara, Geografie, XI-XII, pp. 31–46.

Vuia, F., 2003. Utilizarea hărţilor digitale în studiul unor circuri glaciare din Munţii Parâng. Revista de Geomorfologie 4–5, 25–34.

Wohlfarth, B., Hannon, G., Feurdean, A., Ghergari, L., Onac, B.P., Possnert, G., 2001. Reconstruction of climatic and environmental changes in NW Romania during the early part of the last deglaciation (15,000–13,600 cal years BP). Quaternary Science Reviews 20, 1897–1914.

Woodward, J.C., Macklin, M.G., Smith, G.R., 2004. Pleistocene glaciation in the mountains of Greece. Europe. In: Ehlers, J., Gibbard, P. (Eds.), Quaternary Glaciations – Extent and Chronology, vol. 1. Elsevier, Amsterdam, pp. 271–294.

Xu, S., Dougans, A.B., Freeman, S.P.H.T., Schnabel, C., Wilcken, K.M., 2010. Improved ^{10}Be and ^{26}Al-AMS with a 5 MV spectrometer. Nuclear Instruments and Methods in Physics Research B268, 736–738.

Zahno, C., Akçar, N., Yavuz, V., Kubik, P.W., Schlüchter, C., 2010. Chronology of Late Pleistocene glacier variations at the Uludağ Mountain, NW Turkey. Quaternary Science Reviews 29, 1173–1187.