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First cosmogenic geochronology from the Lesser Caucasus: Late Pleistocene glaciation and rock glacier development in the Karçal Valley, NE Turkey

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A R T I C L E I N F O

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A B S T R A C T

Evidence of widespread alpine glaciations during the Late Pleistocene in mid-latitudes has long attracted attention of researchers. However, there were no studies that contain absolute ages in the Lesser Caucasus despite the fact that it is one of the major glaciated regions in Eurasia. Here, we present first cosmogenic ^{36}Cl surface exposure ages from the Karçal Mountains (41.24° N, 42.06° E, 3431 m a.s.l., above sea level) which is located in the most western part of the Lesser Caucasus in the northeastern Anatolia. In the Karçal Mountains, there are numerous valleys that have experienced significant glaciations since Late Pleistocene. We have investigated one of the largest valleys, the east-facing Karçal Valley, that hosts even a small (2926 m a.s.l.) recent glacier located at above 3000 m a.s.l. Fossil and recent rock glaciers along with lateral and recessional moraines exist in the valley. We conducted the study in two stages. First, we mapped the geomorphological units in the Karçal Valley in detail based on our field works and aerial photography. Later, we collected 10 rock samples from the fossil rock glacier and recessional moraines for cosmogenic ^{36}Cl surface exposure dating. The results outline a glacial chronology that is typical of the Last Glacial Maximum. Although the maximum extent and timing of the glaciation is not exactly known as lateral and terminal moraines were not suitable for sampling, recessional moraines indicate that the Karçal Valley palaeoglacier deglaciation started at least 19.9 ± 1.2 ka ago. Fossil rock glacier samples were dated to 15.7 ± 1.3 ka. These quantitative results are first in the Lesser Caucasus and compatible with previous ages obtained from other valleys in the nearby Eastern Black Sea region, Anatolian and some of the European Mountains.

1. Introduction

The high altitude mountains of Anatolia witnessed widespread glaciations during the Late Pleistocene (126–11.7 ka) (Fig. 1). Although most of these mountains have no recent glaciers, the geomorphological evidences show widespread palaeoglacier activity in the past. This is true especially during the Last Glacial Maximum (LGM), which refers to the peak in global ice volume during the last glacial cycle between 23.000 and 14.000 ^{14}C ka BP, with a mid-point at 18.000 ^{14}C ka BP (Shackleton et al., 1977; Hughes and Gibbard, 2015). This is documented both from the marine oxygen isotope record and from the global sea levels in corals (Mix et al., 2001). On the other hand continental data examined by Clark et al. (2009) indicated that glaciers reached their maximum positions between 26.5 and 20/19 ka, with rapid deglaciation occurring soon after.

In the beginning of the mid-19th century, evidence of actual and palaeoglaciers in Anatolia has been mentioned in the reports of many European researchers (Ainsworth, 1842; Palgrave, 1872; Krenek, 1932; Bobek, 1940; Messerli, 1967; Birman, 1968). Since the mid-20th century, Turkish scientists have also begun to publish their studies about glaciation in Anatolia (Erinç, 1944, 1952; Bilgin, 1972; Kurter, 1991; Doğu et al., 1993; Çiner, 2003, 2004; Sarıkaya, 2012a; Sarıkaya and Çiner, 2015). Over the last 10 years, the glacial chronologies of Anatolia have been refined with cosmogenic surface exposure dating studies (e.g., Akçar et al., 2007, 2008; Çiner et al., 2015; Sarıkaya et al., 2008; Sarıkaya and Çiner, 2017 and

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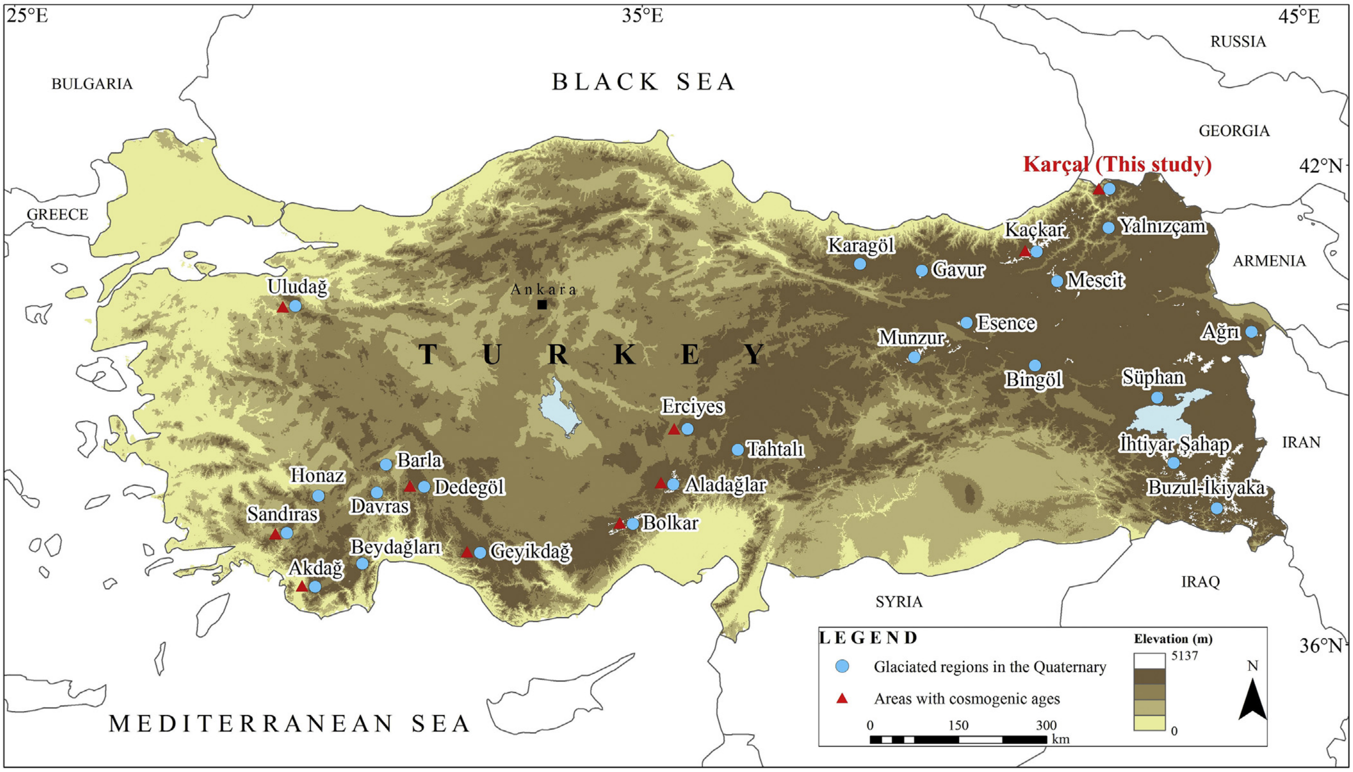

Fig. 1. The distribution of high mountainous areas that experienced palaeoglaciation and locations where cosmogenic surface exposure dating was applied in Turkey.

references therein; Reber et al., 2014; Zahno et al., 2009; Zreda et al., 2011) to the point that Turkey is now considered to have one the best-dated records of its kind in the world (Hughes and Woodward, 2017).

Here, we present our new findings on the Karçal Mountains, which is located in the most northwestern part of the Lesser Caucasus Mountains that runs parallel to the Greater Caucasus Mountains, aligned W-NW to E-SE, ~100 km to its south. This mountainous mass is one of the regions with the clearest evidence of Late Pleistocene glaciation in northeastern Turkey. However, it has been discussed in few studies probably because it is less accessible than the nearby Eastern Black Sea Mountains. Late Pleistocene glaciation in the Karçal Mountains first attracted the attention of Rickmer-Rickmers (1900, 1934), who reported large glaciers, one of the three in the Eastern Black Sea region. Later, Gürgen and Yeşilyurt (2012) reported five cirque glaciers in the Karçal Mountains. Recently, Dede et al. (2015) classified rock glaciers in five valleys (Çamdalı, Karçal, Sakız, Yamukdiken and Ziyaret valleys) on the Karçal Mountains, taking into account their appearance and formation mechanisms based on permafrost creep (Wahrhaftig and Cox, 1959) and ice cored rock glacier (Whalley and Martin, 1992) classification system.

The aim of this study is to provide detailed geomorphological evidence of the Late Pleistocene glaciations in the Karçal Valley, one of the largest valleys in the region, and elucidate the first glacial chronology of the mountain range using cosmogenic ^{36}Cl surface exposure ages obtained from recessional moraines and a fossil rock glacier. Later, we compared our results with nearby Lesser and Greater Caucasus Mountains in Georgia and Armenia, Eastern Black Sea Mountains in Turkey and European Mountains in general. Finally, we presented the current status on surface exposure dated moraine ages in whole Turkish glaciated mountains in order to understand the extent and timing of Last Glacial Maximum (LGM) glaciations in this part of the world.

2. Physical geography and geologic settings

The Karçal Mountains are located in the western part of the Şavşat district in northeastern Anatolia. The deep fluvial tributaries of the Çoruh River separate the Karçal Mountains from the Eastern Black Sea Mountains. The highest peak of the Karçal Mountains is ~40 km away from the Black Sea coast and ~15 km from the Georgian border (Fig. 2).

Within the study area, Karçal Peak (3431 m a.s.l.) is the highest point and the valley floor where Çermik Creek flows, is the lowest point (1500 m a.s.l.). Çermik Creek and its tributaries flow into the Çoruh River and incise deep valleys into the eastern slopes of the mountain. The Karçal Valley, named after the mountain, is one of the major ones, and runs about 5 km E-W on the eastern side of mountain. The width of the valley is 500 m upstream, and reduces to less than 250 m downstream. The slope of the floor of the Karçal Valley varies between 15 and 30%.

The geology of the Karçal Mountains is mainly composed of Cretaceous volcanic rocks, limestones and Eocene volcanics consisting of andesitic-basaltic lavas and pyroclastic rocks (Keskin, 2013). Main lithologies within the Karçal Valley are made up of diorite, hornblende, dacite and rhyolite (Yılmaz et al., 1997). These units are partly covered by Quaternary moraines and alluvium along the valley floor (Fig. 3).

Since there is no nearby meteorological station in the study area, we used the climate data set provided by Hijmans et al. (2005) at www.worldclim.org with 1 km^2 spatial resolution to estimate prevailing climatic conditions. This data indicate that the average annual temperature is around 0 °C in the study area. While average summer temperatures do not exceed 10 °C, winter averages are below −10 °C. Depending on the altitude, precipitation values are about twice as those at Artvin station, which is the nearest meteorological station located at 628 m a.s.l. In the study area, more than half of the precipitation (55%), which is approximately 1000 mm

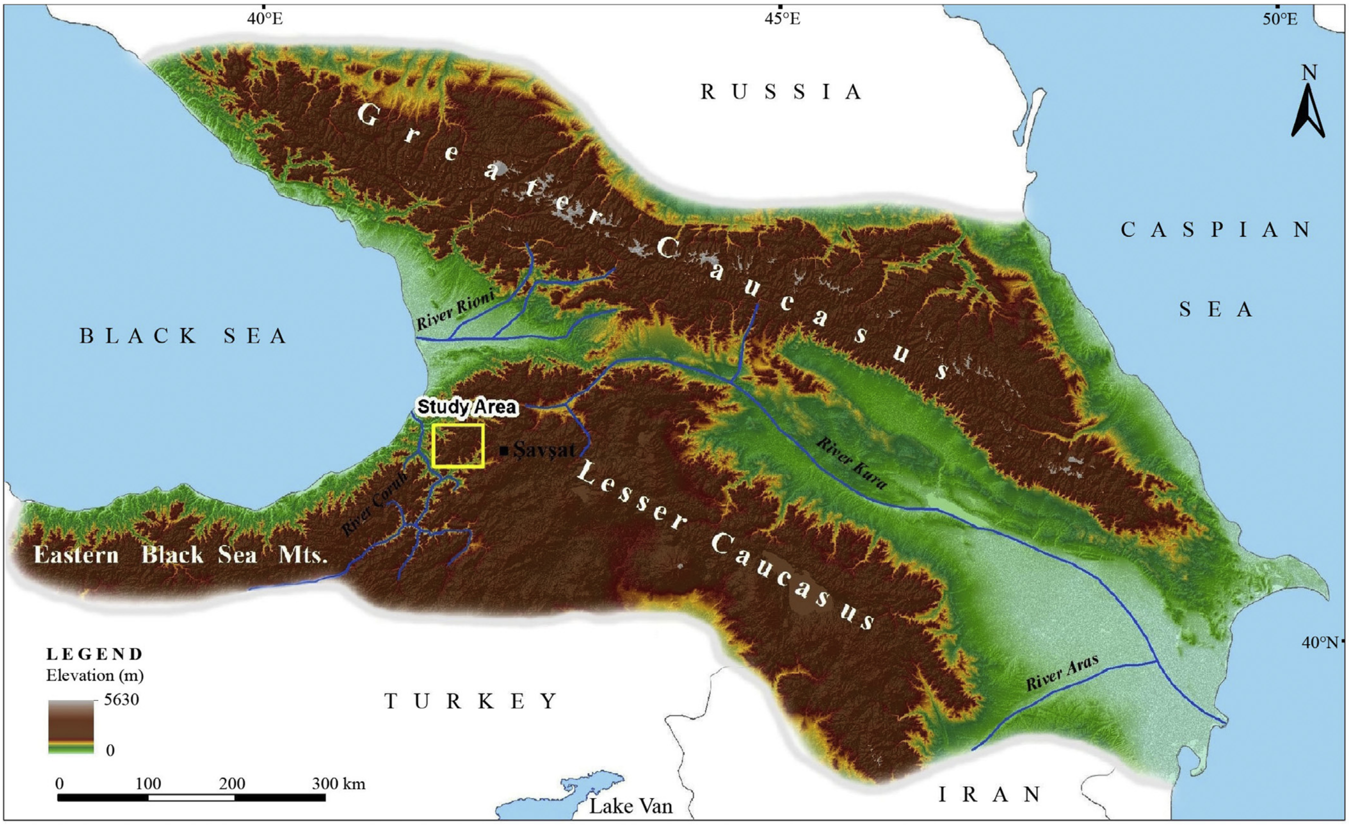

Fig. 2. Digital elevation map of the Caucasus and the location of the Karçal Mountains.

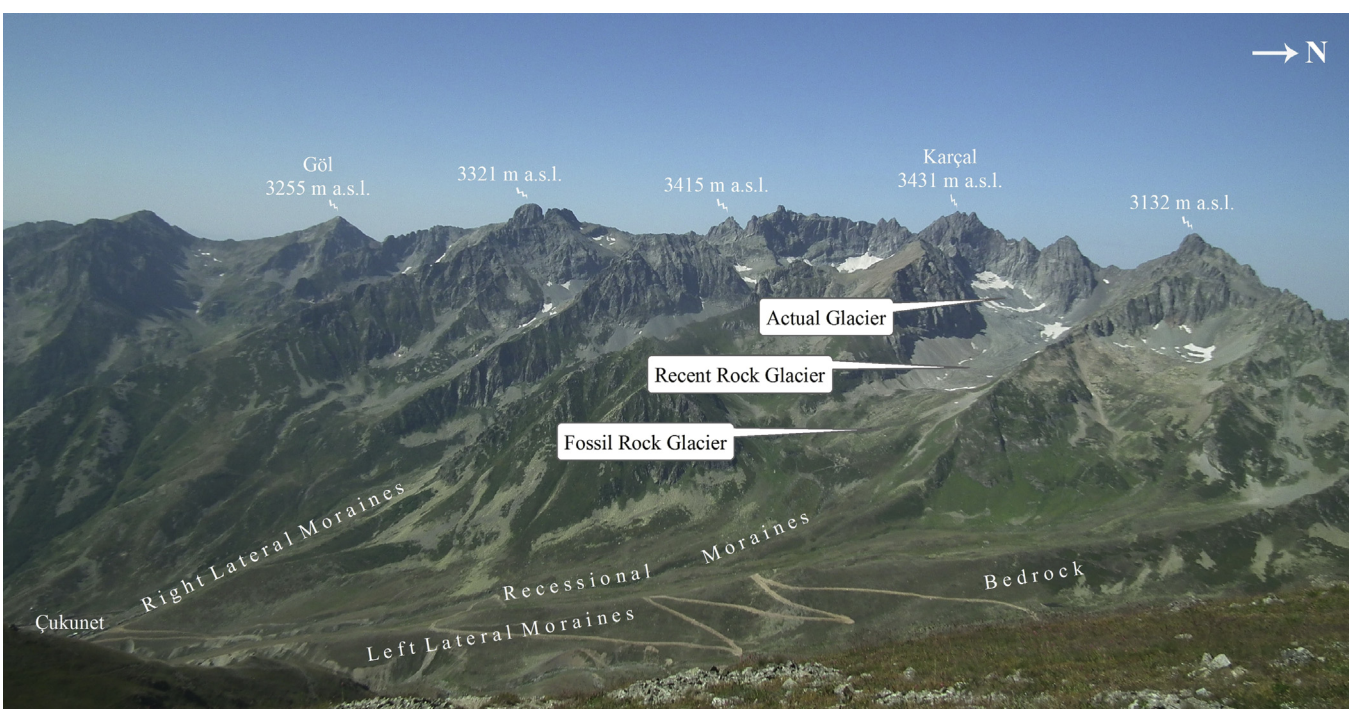

Fig. 3. General view of the Karçal Mountains and the Karçal Valley (looking to west).

per year, falls in the summer months. This shows the continentality effect on the distribution of precipitation. A tundra climate with very short summers prevails in the higher parts of the Karçal Mountains, following Köppen's climate classification (Erlat, 2014).

3. Glacial geomorphology

The Karçal Valley has 5 cirques. There is also an actual glacier within the northern cirque of Karçal Peak, located at above 3000 m a.s.l. Other cirques without glaciers are located between 2830 and 3030 m a.s.l. According to our field observations in August 2015, the lower limit of the actual glacier (i.e. the tongue) was at 2926 m a.s.l. The fact that, the actual glacier is covered by snow during most of the year and high peaks in the south are among the reasons for maintaining its existence until today. The actual glacier is between 3230 m and 2926 m a.s.l. elevations and is adhered to the cirque walls in the northeast. At 3020 m a.s.l. elevation, it is divided into two parts. Its length is ~400 m, with a width of 200 m and a thickness of more than 20 m in places. Further below a recent rock glacier that descends down to 2730 m a.s.l. is observed. The boundary between the actual glacier and the rock glacier is not clear in the field (Figs. 4A and 5).

Recent rock glacier is located at the bottom of the valley between 2935 m and 2730 m a.s.l. It has a length of approximately 1200 m, a width of 250 m and a thickness of 40 m near its tongue. The recent rock glacier has distinct lobes and consists of debris ranging between 10 cm and 200 cm in size. It is covered with lichens and alpine vegetation near its tongue (Fig. 4A).

Fossil rock glacier begins immediately in front of the recent rock glacier and continues until a threshold bedrock ridge on which roches moutonnées and moraines are found to gether at around 2540 m a.s.l. The fossil rock glacier is shaped as longitudinal lobes to be parallel to the direction of the extension of the valley. The lobes consist of ridges 2–3 m high with 1–2 m deep depressions in between. Angular blocks that make up the fossil rock glacier are completely covered with alpine vegetation. Block sizes range from 2 to 10 m. The threshold divides the glacial valley into two parts by a sharp escarpment, which is ~50 m high (Fig. 4A–B). The fossil rock glacier in the Karçal Valley starts immediately in front of the recent rock glacier that ends at 2540 m a.s.l.

There are right and left lateral moraines between 2465 m and 2150 m a.s.l. in the Karçal Valley. The right lateral moraine, 20 m high and 500 m long, extends to the Çukunet settlement starting from 2317 m a.s.l. The left lateral moraine, 25 m high and 100 m long, starts at 2465 m a.s.l. where it is in contact with the bedrock and goes down till 2150 m a.s.l. The fossil rock glacier and recessional moraines are intertwined from the threshold down to 2430 m a.s.l. (Fig. 4C–D).

Several recessional moraines continuing down to 2150 m a.s.l. extend in E-W direction in the form of ridges across the valley. They have a height of approximately 4–5 m with a width of 2–3 m (Fig. 4C). The moraines consist of angular blocks and were deposited during the recession of the glacier in the direction parallel to the extension of the valley. The moraines below this altitude were largely modified by erosion and transported along steep slopes (Fig. 4D).

Today, the permanent snowline is located at 3250 m a.s.l. During the LGM, the snowline was at 2500 m a.s.l. (Messerli, 1967). Hence, glaciers would have developed within a number of valleys oriented E-W during the LGM. Field evidences indicate that the length of the Karçal palaeoglacier was about 5 km and its tongue most likely reached around 2000 m a.s.l. However, it is impossible to determine its exact altitude because of severe ongoing erosion, avalanches and slope debris, which greatly deformed the original structure of the terminal moraines (Fig. 4D).

4. Methodology

To understand the Late Pleistocene glacial chronology of the Karçal Valley, rock samples from fossil rock glacier and recessional

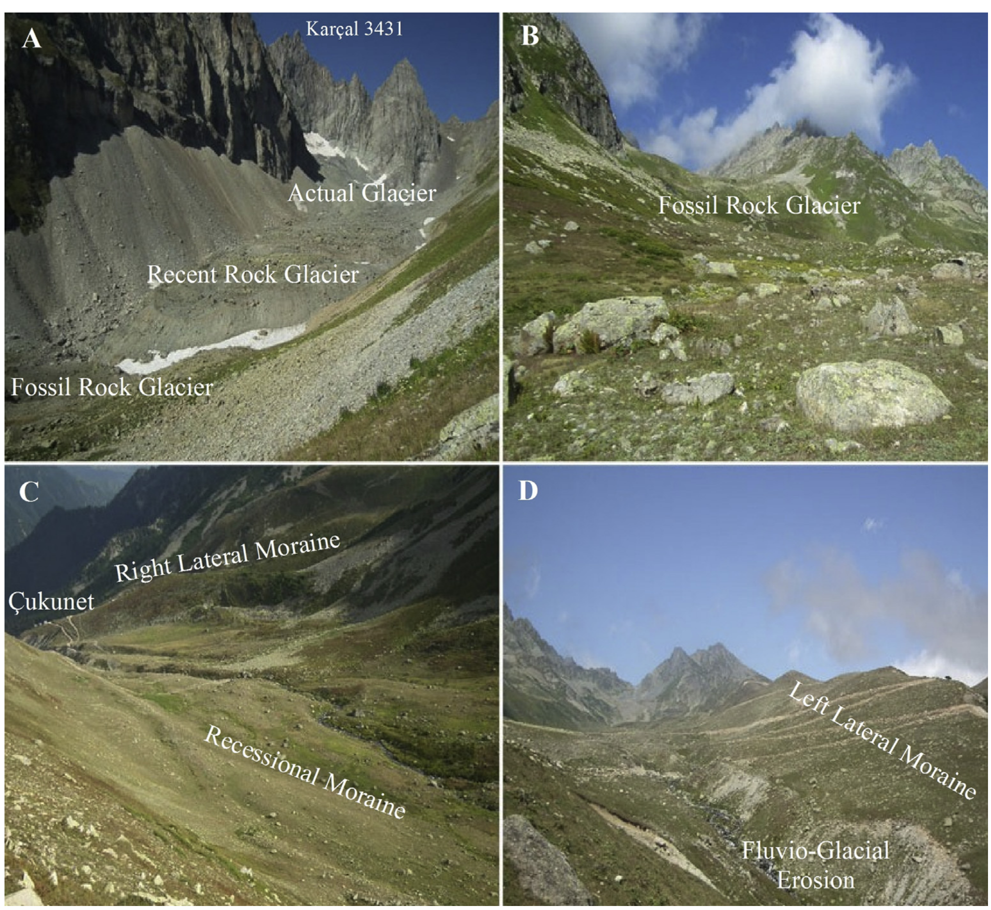

Fig. 4. Field pictures of the Karçal glacial valley. (A) Cirque area, (B) fossil rock glacier, (C) recessional moraine area, and (D) the lowest part of the valley.

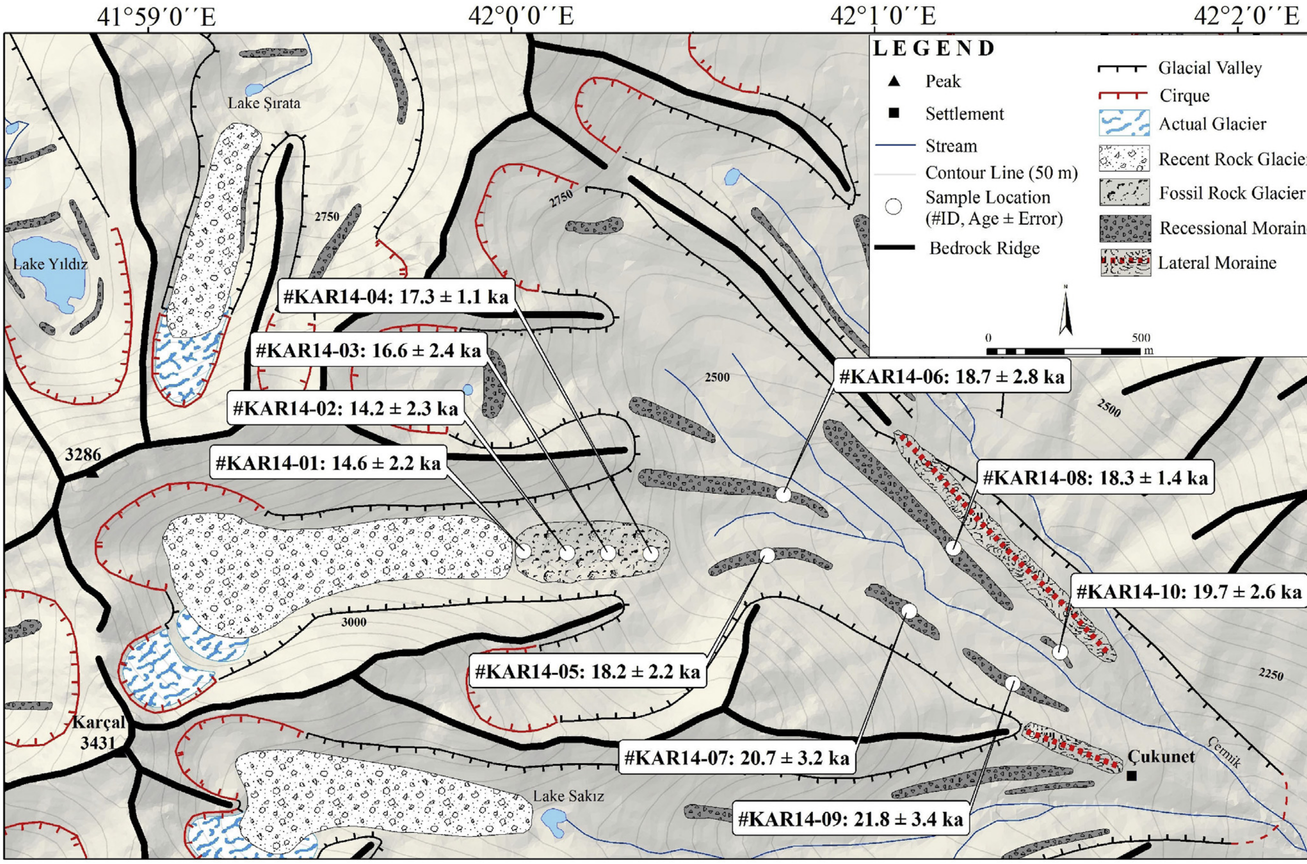

Fig. 5. Glacial geomorphology map of the Karçal Valley. Sample locations with cosmogenic surface ages are indicated.

moraine crests were taken and analysed in laboratories in Turkey, Canada and France. Cosmogenic surface exposure dating was used to assign the glacial chronologies. This method simply estimates when a rock or sediment was exposed to cosmic radiation (Dunai, 2010).

Cosmic rays produced during the galactic explosions and the sun's radioactivity arrives to the Earth in all directions with high energies (Dunai, 2010). These cosmic rays pass through the atmosphere and form new isotopes that attenuate 2–3 m into the surface of rocks. The most commonly used cosmogenic isotopes are ^{10}Be, ^{26}Al, ^{14}C, ^{36}Cl and ^{41}Ca which can be measured with Accelerated Mass Spectrometry (AMS), in very low quantities. The date when a surface exposed to cosmic radiation can be determined based on cosmogenic isotope concentrations in samples taken from the surface. Because the production rate of this newly produced isotopes are known, one can calculate the age of the surface (Lal, 1991; Evans et al., 1997; Stone, 2000; Sarıkaya, 2012b).

4.1. Fieldwork and sampling

The fieldwork in the Karçal Mountains was conducted during the summer months between 2013 and 2015. Geomorphological units were mapped and 4 samples from fossil rock glacier and 6 samples from recessional moraines were taken. While taking samples, the largest *in-situ* buried blocks located on the crests of moraines were preferred. Samples were taken with a hammer and chisel. Rocks were chipped with a thicknesses ranging from 0.5 cm to 1.2 cm from the exposed surfaces so that each sample would be about 700–800 g. The altitude and coordinates of each sample location were taken with a hand-held GPS (Garmin Dakota) (Table 1). The topographic horizon angles for each location were measured with an inclinometer to correct for the topographical shielding (Gosse and Phillips, 2001). Snow corrections were made by the predictions of each month's snowfalls based on the interpolated climate date provided by Hijmans et al. (2005), (www.worldclim.org). The CRONUS Web Calculator v.2 (www.cronuscalculators.nmt.edu) was used to calculate the exposure ages of our samples (Marrero, 2012).

4.2. Sample preparation

Samples were prepared in Kozmo-Lab in Istanbul, Turkey. Lithological properties were analysed using thin sections prepared at the Istanbul Technical University (Table 2). Samples were first crushed and ground and then sieved to 0.25–1 mm. Later, samples were leached with diluted HNO_3 (10%) overnight. Next, about 5 g of leached samples were digested in pressure chambers (Parr #4748) with 40 ml HF (40%) and 5 ml HNO_3 (65%). The pressure chambers were placed in an oven at 130 °C for 6 hours to increase the digestion efficiently. Then, ^{35}Cl (99.7%) enriched Aldrich spike added to samples and total chlorine was separated from the rock matrix by adding $AgNO_3$. Several steps of $BaSO_4$ precipitation were applied to remove the ^{36}S isobar effect. The final AgCl precipitates were sent to the AMS Laboratory in Centre Européen de Recherche et d'Enseignement des Géosciences de l'Environnement (CEREGE) in France for isotope ratio measurements. Major and minor elemental compositions of the samples were measured in the Acme Laboratory in Canada (Table 2). The details of sample preparation procedures were described in Sarıkaya (2009).

5. Results

Ten boulders were sampled from recessional moraine crests and

Table 1
Field parameters of the cosmogenic samples.

Sample name	Latitude (WGS84) °N (DD)	Longitude (WGS84) °E (DD)	Elevation (m)	Boulder dimension (m)	Sample thickness (cm)	Topography correction factor (-)	Snow correction factor (-)
KAR14-01	41.3548	42.0026	2712	1 × 2 × 2	0.5	0.9949	0.8906
KAR14-02	41.3546	42.0056	2656	1 × 2 × 2	1	0.9953	0.8906
KAR14-03	41.3548	42.0075	2617	1.5 × 2 × 1.5	0.6	0.9886	0.9204
KAR14-04	41.3549	42.0087	2564	0.5 × 1 × 1	0.5	0.9939	0.8868
KAR14-05	41.3550	42.0138	2433	1 × 2 × 2	1	0.9905	0.9099
KAR14-06	41.3567	42.0144	2425	3 × 4 × 3	0.5	0.9899	0.9985
KAR14-07	41.3540	42.0177	2371	3 × 4 × 3	1.2	0.9967	0.9971
KAR14-08	41.3564	42.0201	2377	1 × 2 × 2	0.8	0.9940	0.9061
KAR14-09	41.3503	42.0265	2233	1.5 × 2 × 1.5	1	0.9952	0.9300
KAR14-10	41.3506	42.0285	2180	2 × 2 × 2	0.6	0.9928	0.9564

Table 2
Geochemical and cosmogenic isotope analysis data of the samples.

Sample name	Lithology	Major elements (wt. %)										
		Al_2O_3	CaO	Fe_2O_3	K_2O	MgO	MnO	Na_2O	P_2O_5	SiO_2	TiO	CO_2
KAR14-01	Diorite	17.82	5.73	6.94	0.54	3.43	0.19	3.53	0.19	57.10	0.65	3.70
KAR14-02	Andesite	18.05	5.25	6.41	2.98	2.64	0.13	4.14	0.04	57.39	0.55	2.20
KAR14-03	Diorite	18.47	6.09	6.57	1.80	3.13	0.13	4.09	0.18	55.77	0.59	3.00
KAR14-04	Diorite	19.48	8.90	7.67	0.48	4.97	0.18	3.56	0.07	50.09	0.78	3.60
KAR14-05	Gabbro	17.19	5.65	6.11	2.80	3.38	0.11	4.14	0.16	58.64	1.04	0.50
KAR14-06	Diorite	14.89	7.89	5.37	5.65	0.73	0.06	1.64	0.14	59.02	0.39	4.10
KAR14-07	Diorite	18.15	5.13	5.54	2.55	2.07	0.14	3.91	0.05	58.54	0.51	3.20
KAR14-08	Basalt	18.17	5.55	6.50	2.04	3.13	0.12	3.68	0.09	56.05	0.60	3.90
KAR14-09	Diorite	17.35	6.54	6.49	0.91	4.40	0.16	2.81	0.15	55.98	0.64	4.30
KAR14-10	Tuffite	18.05	4.32	4.22	1.26	2.32	0.05	3.89	0.07	62.76	0.43	2.40

Sample name	Surface[a]	Trace elements (ppm)					
		Sm	Gd	U	Th	Cl	$^{36}Cl/Cl$ (10^{-15}) (-)
KAR14-01	FRG	3.37	3.44	0.90	3.80	31.7 ± 3.0	736 ± 11
KAR14-02	FRG	3.03	3.05	0.80	4.40	15.4 ± 1.5	1887 ± 29
KAR14-03	FRG	3.23	3.39	1.00	3.90	29.9 ± 2.8	1105 ± 14
KAR14-04	FRG	3.06	3.21	0.70	2.90	16.4 ± 1.6	1621 ± 23
KAR14-05	RM	2.18	2.28	1.10	5.60	144.0 ± 13.2	439 ± 4
KAR14-06	RM	4.38	4.48	2.30	10.40	40.6 ± 3.8	1651 ± 18
KAR14-07	RM	3.41	3.17	0.80	3.40	12.0 ± 1.2	2888 ± 38
KAR14-08	RM	3.08	3.07	1.10	3.80	19.6 ± 1.9	1380 ± 23
KAR14-09	RM	3.18	3.19	1.00	2.50	22.5 ± 2.2	1232 ± 16
KAR14-10	RM	2.17	2.21	2.30	7.00	36.7 ± 3.4	742 ± 9

[a] **FRG:** Fossil rock glacier boulders, **RM:** Recessional moraine boulders.

fossil rock glacier in the Karçal Valley (Fig. 5). We have reported two different surface exposure ages for each sample with and without erosion correction (Table 3). The erosion corrected ages were calculated with an erosion rate of 5 mm/ka, as proposed in similar lithologies elsewhere (e.g., Palacios et al., 2011). Erosion does not affect the ages by much, i.e. zero erosion and erosion produced ages are within their error margins. This rate is based on our field observations of physical erosion marks on boulder surfaces ranging in depth from a few mm to maximum 10 cm. According to the ages range between 18.2 ± 2.2 ka to 21.8 ± 3.4 ka for the recessional moraines and 14.2 ± 2.3 ka to 17.3 ± 1.1 ka for the fossil rock glacier samples. All age uncertainties were given in 1-sigma level. The samples were grouped according to their positions in the landforms and presented in Table 3.

The erosion corrected ages from each landform indicate two distinct glaciation periods: LGM (23–18 ka) represented by recessional moraines and the Late-glacial (18–13 ka) represented by fossil rock glaciers (Fig. 5).

5.1. Recessional moraines

Although right and left lateral moraines exist, they were not suitable for sampling as their surface is partly disturbed by a new road construction. Therefore six boulder samples were collected from the recessional moraines between the altitudes of 2433 m and 2180 m a.s.l. (Figs. 5 and 6). The right recessional moraine samples (KAR14-05, KAR14-07 and KAR14-09) yielded boulder ages of 18.2 ± 2.2 ka, 20.7 ± 3.2 ka and 21.8 ± 3.4 ka, collected from 2433, 2371, and 2233 m a.s.l., respectively. On the other hand, the left recessional moraine samples (KAR14-06, KAR14-08 and KAR14-10) yielded boulder ages of 18.7 ± 2.8 ka, 18.3 ± 1.4 ka and 19.7 ± 2.6 ka and were taken from 2425, 2377 and 2180 m a.s.l., respectively (Table 3). The weighted average age of all samples from the recessional moraines is 19.9 ± 1.2 ka.

5.2. Fossil rock glacier boulders

Four boulder samples were taken along a line in the upper part of the threshold that divides the Karçal Valley (Figs. 5, 7 and 8). The first sample KAR14-01 was collected at 2712 m a.s.l. and dated to 14.6 ± 2.2 ka ago. The other samples KAR14-02, KAR14-03 and KAR14-04 were collected from 2556, 2617 and 2564 m a.s.l. and dated to 14.2 ± 2.3, 16.6 ± 2.4, and 17.3 ± 1.1 ka ago, respectively (Table 3). The weighted average age of all fossil rock glacier samples

Table 3
Cosmogenic surface ages from the Karçal Valley.

Sample name	Surface[a]	^{36}Cl (measured) (10^4 atoms g^{-1} rock)	Depth average total production rate (atoms g^{-1} rock a^{-1})	Surface exposure ages[b]		Landform age[c] (ka)
				Without erosion correction (ka)	Erosion corrected (5 mm ka^{-1})	
KAR14-01	FRG	39.7 ± 3.0	31.6	15.4 ± 2.5	14.6 ± 2.2	**15.7 ± 1.3**
KAR14-02	FRG	49.4 ± 3.5	45.1	13.8 ± 2.3	14.2 ± 2.3	
KAR14-03	FRG	56.2 ± 3.3	40.2	16.8 ± 2.5	16.6 ± 2.4	
KAR14-04	FRG	45.2 ± 3.0	33.7	16.7 ± 3.1	17.3 ± 1.1	
KAR14-05	RM	107.4 ± 4.3	62.1	20.2 ± 2.7	18.2 ± 2.2	**19.9 ± 1.2**
KAR14-06	RM	113.8 ± 4.8	69.2	18.3 ± 2.8	18.7 ± 2.8	
KAR14-07	RM	59.1 ± 3.1	32.9	19.8 ± 2.9	20.7 ± 3.2	
KAR14-08	RM	45.9 ± 4.0	31.7	17.5 ± 2.9	18.3 ± 1.4	
KAR14-09	RM	47.0 ± 2.9	24.8	21.8 ± 3.4	21.8 ± 3.4	
KAR14-10	RM	46.2 ± 2.5	24.0	21.2 ± 2.9	19.7 ± 2.6	

[a] **FRG:** Fossil rock glacier, **RM:** Recessional moraine.
[b] The uncertainties of boulder ages were given at the 1 sigma level.
[c] Weighted average of the landform based on 5 mm ka^{-1} erosion corrected boulder ages.

is 15.7 ± 1.3 ka.

6. Discussion

Our results from the Karçal Mountains are important for the glacial chronology in Turkey, as there were very few previous studies and no quantitative ages in the whole Lesser Caucasus Mountains. Here, we discuss the cosmogenic ^{36}Cl surface exposure ages in Karçal Mountains and compare them with nearby Lesser and Greater Caucasus Mountains where even relative age data are scarce. On the other hand, cosmogenic ^{10}Be ages are very well established in the nearby Black Sea Mountains and ^{36}Cl and ^{10}Be ages on several other mountains of Anatolia (Table 4). We also briefly described those mountains together with some major European sites for comparison purposes.

6.1. Karçal Valley chronologies

The moraine deposits and other glacial erosional features in Karçal Valley, which can be traced in the valley bottom, indicate that the Karçal palaeoglacier reached around 2100 m a.s.l. and joined other glaciers from Çukunet Valley. Accordingly, the length of the palaeoglacier in the Karçal Valley between cirques and the junction of two glaciers was at least 5 km. The absence of the terminal moraines in the valley is probably related to the high slope and increased precipitation after the deglaciation. The boulders composing the terminal moraines were removed to the lower parts of the valley with post-glacial erosion.

The recessional moraine boulder ages of 18.2 ± 2.2 ka to 21.8 ± 3.4 ka probably mark the deglaciation times of the glacier after the local LGM. The weighted average age of the recessional moraine samples yields 19.9 ± 1.2 ka (Table 3; Figs. 5 and 9). However, although we do not have quantitative age data, the maximum extent of the local LGM glacier was ~500 m lower and probably several thousands years earlier than the recessional moraines. This widest extension sits in the middle of Greenland Stadial 3 (27.5–23.3 ka), which is notable across Europe and the northern hemisphere for its significant glacier advances. Accordingly, Hughes and Gibbard (2015) argued that the global LGM could be correlated with the Greenland Stadial 3. Although we only have recessional moraine ages (19.9 ± 1.2 ka) in the Karçal Valley, the maximum extent of the glaciation could be tentatively correlated with this Greenland Stadial 3. While the colder and somewhat wetter climate conditions during the local LGM caused the glaciers to expand, the recession might have occurred depending on the increasing drier conditions (Shumilovskikh et al., 2012).

On the other hand, other ages from 14.2 ± 2.3 ka to 17.3 ± 1.1 ka obtained from the fossil rock glacier show an enhancement of the periglacial conditions, which probably occurred during the Lateglacial in the region. The weighted average age of the fossil rock glacier samples is 15.7 ± 1.3 ka (Table 3; Figs. 5 and 9). This age indicates that there was still a glacier present in the area between the rock glaciers and the threshold, but the glacier in the lower part of the threshold rapidly retreated after the local LGM.

6.2. Lesser and Greater Caucasus glaciations

The Georgian and Armenian parts of the Lesser Caucasus also experienced glaciations during the Late Pleistocene. The character of glaciation resembles the central parts of Greater Caucasus, where the altitudes are comparably low (Gobejishvili et al., 2011). In the Lesser Caucasus the Würmian-age firn line was lowest on the ranges located closest to the Black Sea (2200–2300 m a.s.l.) and highest on the ranges situated to the east (2500–2600 m a.s.l.) (Akhalkatsishvili et al., 2003).

On the Georgian part, the largest valley glaciers (4–6 km long) developed on the Samsari-Javakheti Range (3285 m a.s.l.), where the cirque lower threshold was located at 2500–2700 m a.s.l. (Akhalkatsishvili et al., 2003). Although there are no active glaciers today, these volcanic mountains provided favourable conditions for the development of glaciers that descended down to 1450–1500 m a.s.l. (Messager et al., 2013). The presence of moraine fields in the eastern part of Lake Paravani shores (2073 m a.s.l.) on the Samsari-Javakheti Range, which covers volcanic rocks of known age are reported by Messager et al. (2013). The authors proposed that these glacial deposits should be contemporaneous or younger than the MIS-6 (Marine Isotope Stage) glacial period, adding that most of the fresh looking moraines date probably to the Last Glacial period MIS 4-2. On the other hand, in the Syunik volcanic massif, situated in the southern Armenia, glacial landforms such as two terraced moraine deposits at the Shamb area, erratic blocks and fluvioglacial deposits were attributed to colder conditions of the MIS-2, indicating two climatic upheavals (LGM?) (Ollivier et al., 2010).

Although based only on aerial photos, satellite images and few field observations, evidence of actual glaciers and palaeoglaciers in the nearby Greater Caucasus Mountains has been mentioned in several recent articles (Tzereteli, 1974; Gobejishvili et al., 2011; Tielidze et al., 2015a,b,c; Tielidze, 2016; Solomina et al., 2016; Holobâcă, 2016). Because the Caucasus Mountains occur midway between the Alps and the mountains of central Asia, they are of great interest for understanding the dynamics of glaciers formed under both maritime and continental climatic conditions

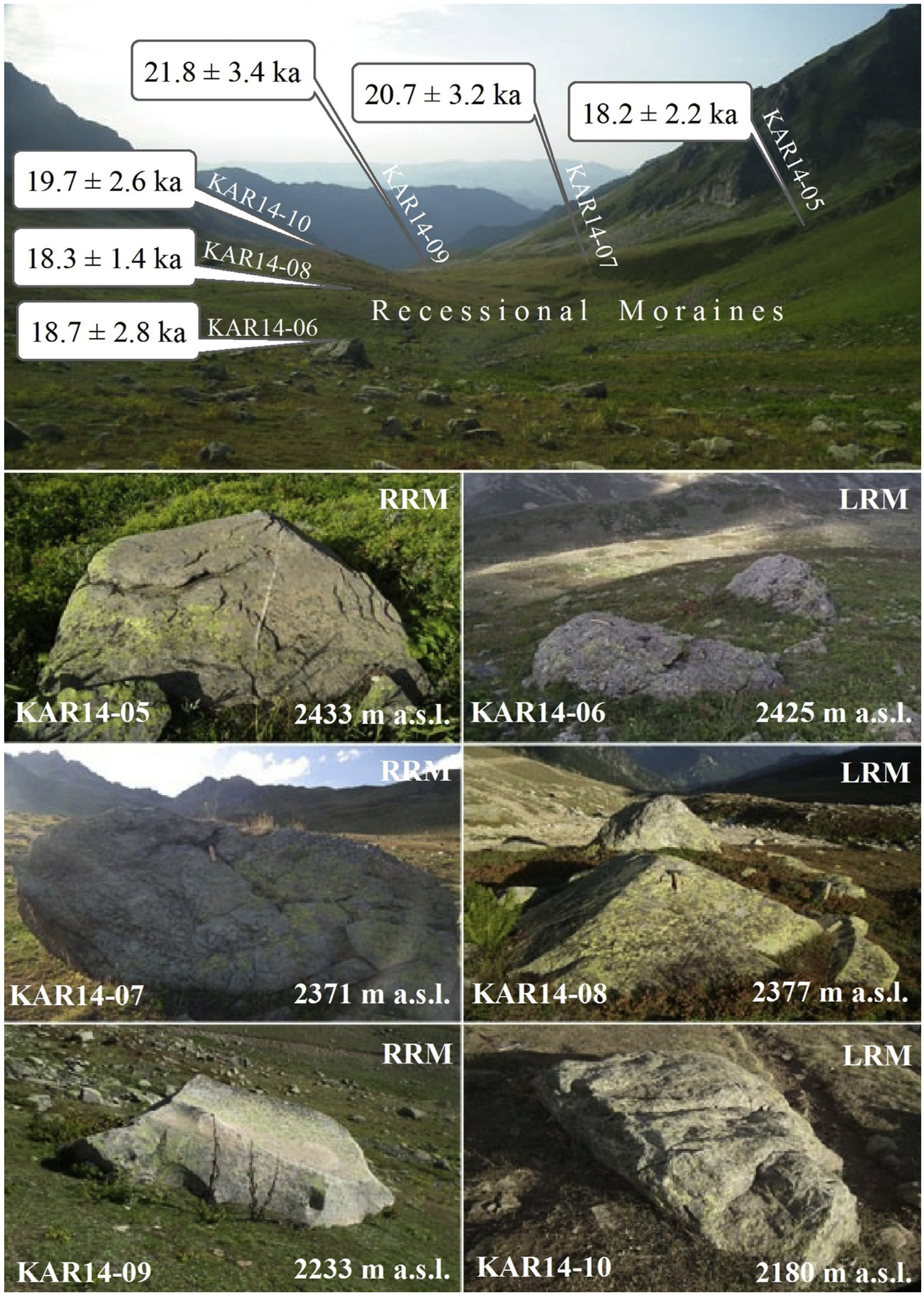

Fig. 6. General view of the recessional moraines of the lower Karçal Valley and the samples with ages. RRM: Right Recessional Moraine, LRM: Left Recessional Moraine.

(Gobejishvili et al., 2011). As a whole, the Late Pleistocene glaciation in the Greater Caucasus consisted of valley glaciers with a few ice caps. Although the north facing glaciers attained lengths up to 50–70 km, outlet glaciers of considerable size (17–35 km length) occurred on both slopes of the central Caucasus with glacier tongues that descended down to ~600 m in places (Gobejishvili et al., 2011).

Active glacial developments in the Greater Caucasus are even better described by several recent studies. For instance, Stokes et al. (2006), used satellite images to obtain the areal fluctuations of 113 glaciers in 1985 and 2000. Their results indicate that 2% of the glaciers advanced, 94% retreated and 4% exhibited no change. In a similar study Shahgedanova et al. (2014) found that 5% of the glaciers in the Elbrus Mountains (out of 498 glaciers) recessed between 1987 and 2010. Although glacier retreat started in the late 1840s in the Elbrus Mountains, retreat has accelerated since 1980 (Holobâcă, 2016). In a review paper concerning the Northern Caucasus, Solomina et al. (2016) also reported the decrease in length of most of the glaciers in the region and pointed out the resemblance between the patterns of climatic and glacier variations with the Alps.

In another study, Stokes et al. (2007), revealed changes in supraglacial debris cover and supra/proglacial lake development associated with glaciers retreat in the central Greater Caucasus

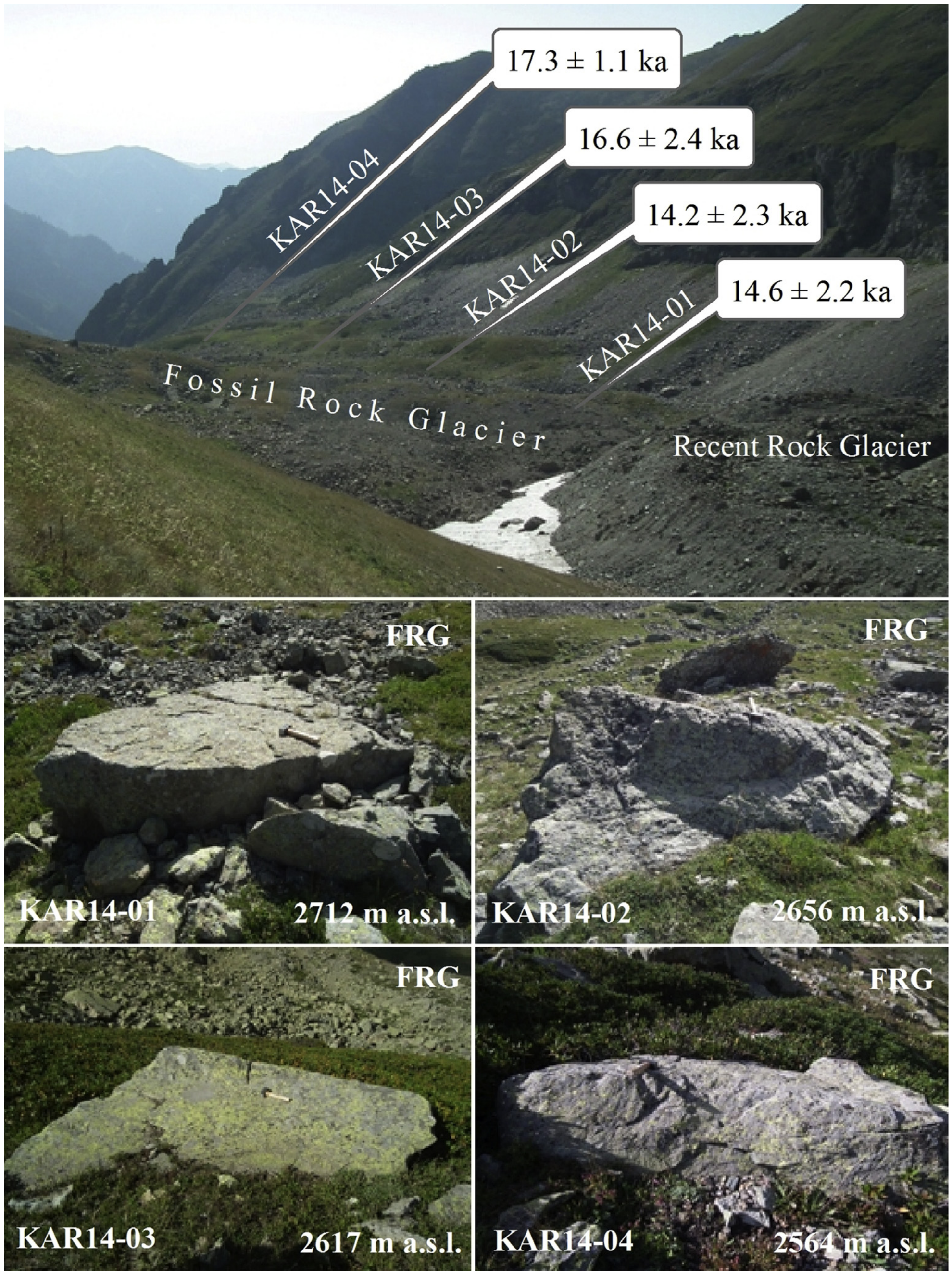

Fig. 7. General view of the fossil rock glacier in the upper Karçal Valley. Sample pictures and ages are presented. FRG: Fossil Rock Glacier.

Mountains in Russia between 1985 and 2000. They concluded that supraglacial debris cover has increased from 5% to 25% and proglacial lakes augmented 57% their surface areas. Tielidze et al. (2015a) estimated 30% recession in the glaciers of Dolra River basin in the Greater Caucasus in 1911–1960 and 2014. Topographical maps, aerial photos and climatic data were used for the glaciers of Mulkhura River basin and the Mestiachala River basin to find out ~20% recession in less than a century (Tielidze et al., 2015b,c).

6.3. Anatolian Mountains glaciations

The number of glacial chronology studies with cosmogenic surface ages in the Anatolian Mountains has increased substantially in the last 10 years. Studies on Turkish Mountains glacial chronologies, based on quantitative data are listed in Table 4.

The glaciations within the Eastern Black Sea Mountains are well-dated and present similarities with ages obtained from Karçal Mountains. Several studies with cosmogenic surface ages from the Kavron, Verçenik and Başyayla glacial valleys (Akçar et al., 2007, 2008; Reber et al., 2014), which are located in approximately 80 km southwest of the Karçal Valley, yield LGM and Late-glacial ages (Table 4). For instance based on 22 dated samples from Kavron Valley, glaciation began no earlier than 27.3 ± 1.7 ka ago and ended 19.8 ± 1.4 ka ago (LGM). The retraction periods in the Kavron Valley were dated to the Late-glacial (17.0 ± 1.1 ka) and Younger Dryas (12.8 ± 1.0 ka) (Akçar et al., 2007). On the other hand cosmogenic surface ages obtained from 19 samples in the Verçenik Valley indicated that the earliest glaciation began 27.5 ± 1.8 ka ago

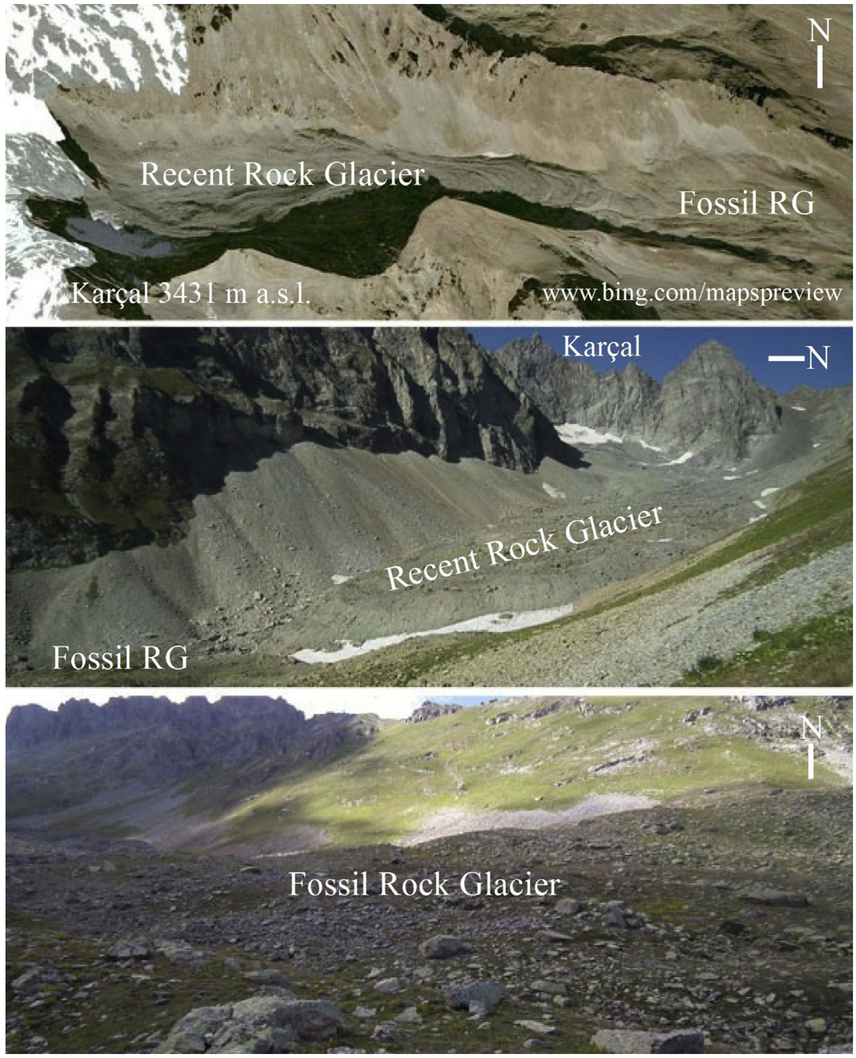

Fig. 8. General view of the recent and fossil rock glaciers.

and ended 20.3 ± 1.4 ka ago (LGM). Late-glacial moraines were dated to 17.2 ± 1.2 ka ago (Akçar et al., 2008). In Başyayla Valley the earliest glaciation began 57.0 ± 3.5 ka ago and the glacier advance continued until 21.2 ± 1.3 ka ago (LGM). The moraines of the Late-glacial age were dated to 17.0 ± 1.0 ka ago (Reber et al., 2014).

Results on the Taurus Mountain Range along the Mediterranean coast of Turkey yield a relatively synchronous development of the glaciers during LGM. For instance in Mount Sandıras (2295 m a.s.l.), the most southwestern glaciated mountain of Turkey, glaciation reached its largest size 22.9 ± 3.3 ka ago in the Kartal Lake Valley and began to retreat 20.6 ± 3.1 ka ago (Sarıkaya et al., 2008). On the other hand in the NW valley of Mount Sandıras deglaciation started in Late-glacial, 17.2 ± 0.6 ka ago (Sarıkaya et al., 2008). Similarly, in Mount Akdağ (3016 m a.s.l.) located also in the Western Taurus, deglaciation started 21.7 ± 1.2 ka ago in the Kuruova Valley, 20.1 ± 0.5 ka ago in the Taşkuzluklu Valley and 19.3 ± 0.8 ka ago in the Karadere Valley (Sarıkaya et al., 2014). Late-glacial ages were also obtained from Kuruova (15.1 ± 0.9 ka), Taşkuzluklu (16.1 ± 0.3 ka) and Karadere (15.7 ± 0.5 ka) valleys (Sarıkaya et al., 2014). Approximately 200 km to the NW of Mount Akdağ, the Muslu Valley in Dedegöl Mountains (2992 m a.s.l.) was also glaciated 29.6 ± 1.9 ka ago and glaciers began to retreat in LGM, 21.5 ± 1.5 ka ago (Zahno et al., 2009). Late-glacial period is dated to 15.2 ± 1.1 ka ago (Zahno et al., 2009).

In the central Taurus Mountains Range, Geyikdağ Mountains (2877 m a.s.l.), also witnessed an extensive glaciation during the LGM. Thanks to the development of piedmont glaciers represented mainly by hummocky moraines, the glaciation reached its maximum extent 19.1 ± 3.4 ka ago and ended 18.0 ± 1.1 ka ago (Çiner et al., 2015). The authors reported that hummocky moraine forming processes with cycles of relief inversion might give rise to up to few thousand years younger apparent boulder (and hence moraine) ages in Geyikdağ. Therefore these ages should be regarded as minimum ages for glacier retreat with true ages close to LGM observed in nearby Taurus Mountains. The Late-glacial (14.0 ± 1.3 ka) is represented by hummocky moraines in the Susam Valley in Geyikdağ (Çiner et al., 2015).

In the other parts of the central Taurus Mountains glacial chronologies are no different. For instance 30 boulder/bedrock samples taken from the Karagöl, Alagöl and Elmalı valleys of the Bolkar Mountains (3524 m a.s.l.) were dated by ^{36}Cl cosmogenic surface method. According to the results, the glaciation in the Karagöl Valley reached its widest extension 25.0 ± 4.1 ka ago and began to retreat 18.9 ± 4.1 ka (LGM) ago.

Late-glacial ages were also obtained from the Karagöl Valley (14.6 ± 2.8 ka) and Alagöl Valley (15.2 ± 1.6 ka) (Çiner and Sarıkaya,

Table 4
Studies and quantitative ages from the Anatolian Mountains.

Study Area	Glacial Valleys	Last Glacial Maximum (LGM) Advance	Last Glacial Maximum (LGM) Deglaciation	Late-glacial (LG)	Younger Dryas (YD)
Uludağ (2542 m a.s.l.)	Ski Area Valley (Zahno et al., 2010)	>24.6 ± 1.5 ka	18.6 ± 1.3 ka	15.2 ± 1.0 ka	
106 samples	Karagöl Valley (Akçar et al., 2014)	>20.4 ± 1.2 ka	18.6 ± 1.2 ka	15.9 ± 1.1 ka	
	Kovuk Valley (Akçar et al., 2015)		18.2 ± 1.3 ka		
Sandıras (2295 m a.s.l.) 12 samples	Kartal Lake Valley (Sarıkaya et al., 2008)	>22.9 ± 3.3 ka	20.6 ± 3.1 ka		
	Northwestern Valley (Sarıkaya et al., 2008)			17.2 ± 0.6 ka	
Akdağ (3016 m a.s.l.) 41 samples	Kuruova Valley (Sarıkaya et al., 2014)	>21.7 ± 1.2 ka		15.1 ± 0.9 ka	
	Taşkuzluklu Valley (Sarıkaya et al., 2014)	>20.1 ± 0.5 ka	19.0 ± 0.4 ka	16.1 ± 0.3 ka	
	Karadere Valley (Sarıkaya et al., 2014)	>19.3 ± 0.8 ka	18.4 ± 0.3 ka	15.7 ± 0.5 ka	
Dedegöl (2992 m a.s.l.) 25 samples	Muslu Valley (Zahno et al., 2009)	>29.6 ± 1.9 ka	21.5 ± 1.5 ka	15.2 ± 1.1 ka	
Geyikdağ (2877 m a.s.l.)	Susam Valley (Çiner et al., 2015)			13.4 ± 1.5 ka	
34 samples	Namaras Valley (Çiner et al., 2015)	>19.1 ± 3.4 ka	18.0 ± 1.1 ka		11.6 ± 2.7 ka
Bolkar (3524 m a.s.l.) 30 samples	Karagöl Valley (Çiner and Sarıkaya, 2017)	>25.0 ± 4.1 ka	18.9 ± 3.3 ka	14.6 ± 2.8 ka	
	Alagöl Valley (Çiner and Sarıkaya, 2017)			15.2 ± 1.6 ka	
	Elmalı Valley (Çiner and Sarıkaya, 2017)				12.6 ± 2.3 ka
Aladağlar (3575 m a.s.l.) 22 samples	Hacer Valley (Zreda et al., 2011)			14.0 ± 1.5 ka	12.0 ± 1.8 ka
Erciyes (3917 m a.s.l.)	Aksu Valley (Sarıkaya et al., 2009)	>21.4 ± 2.9 ka	20.7 ± 2.2 ka	15.6 ± 1.9 ka	
44 samples	Üçker Valley (Sarıkaya et al., 2009)	>23.2 ± 2.1 ka	18.5 ± 2.7 ka	15.2 ± 2.1 ka	
Eastern Black Sea	Kavron Valley (Akçar et al., 2007)	>27.3 ± 1.7 ka	19.8 ± 1.4 ka	17.0 ± 1.1 ka	12.8 ± 1.0 ka
Mountains (3932 m a.s.l.)	Verçenik Valley (Akçar et al., 2008)	>27.5 ± 1.8 ka	20.3 ± 1.4 ka	17.2 ± 1.2 ka	
83 samples	Başyayla Valley (Reber et al., 2014)	>24.8 ± 1.4 ka	21.2 ± 1.3 ka	17.0 ± 1.0 ka	
Karçal (3431 m a.s.l.) 10 samples	Karçal Valley (**This study**)	>21.8 ± 3.4 ka	18.2 ± 2.2 ka	14.2 ± 2.3 ka	

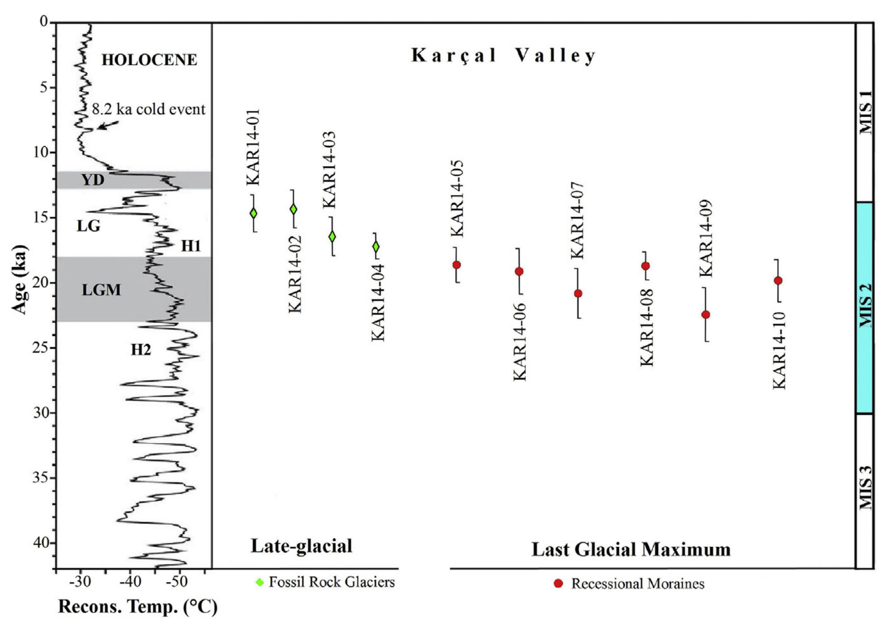

Fig. 9. Reconstructed air temperatures from the GISP 2 ice core in Greenland (Alley, 2000) and cosmogenic surface exposure ages from the Karçal Valley (MIS: Marine Isotope Stages, H: Heinrich events, LGM: Last Glacial Maximum, LG: Late-glacial, YD: Younger Dryas).

2017). Aladağlar (3575 m a.s.l.) is another central Taurus Mountain, where 7 successive terminal moraines were dated with 22 samples. Recalculated cosmogenic surface ages indicate a deglaciation that started in Late-glacial (14.0 ± 1.5 ka) and ended at onset of Holocene including the Younger Dryas (Zreda et al., 2011; Sarıkaya and Çiner, 2017).

In central Anatolia, near Cappadocia, 44 moraine boulder samples taken from the Aksu and Üçker valleys on Erciyes Volcano (3917 m a.s.l.) were also dated. In the Aksu Valley, the glaciation developed during LGM 21.4 ± 2.9 ka ago and ended 20.7 ± 2.2 ka ago. In the Üçker Valley, the glaciation reached its largest size 23.2 ± 2.1 ka ago and began to retreat 18.5 ± 2.7 ka ago. The ages of the Late-glacial are 15.6 ± 1.9 ka in the Aksu Valley and 15.2 ± 2.1 ka in the Üçker Valley (Sarıkaya et al., 2009). Hundred-six rock samples taken from Mount Uludağ (2542 m a.s.l.), located in northwestern Anatolia, indicate that the glaciation developed 24.6 ± 1.5 ka ago in the Ski area and 20.4 ± 1.2 ka ago in the Karagöl Valley, and was terminated 18.6 ± 1.2 ka (LGM) ago. In the region, Late-glacial ages were determined as 15.2 ± 1.0 ka in the Ski area and as 15.9 ± 1.1 ka in the Karagöl Valley (Zahno et al., 2010; Akçar et al., 2014, 2015).

6.4. European Mountains glaciations

Several European mountains (e.g., Alps, Apennines, Pyrenees and Carpathians) were also glaciated during the Late Pleistocene (Hughes et al., 2006, 2013). Today, thanks to the increasing number of publications that use cosmogenic surface exposure data there is a wealth of information concerning the glacial chronologies in Europe. Recent volumes by Palacios and García-Ruiz (2015), devoted to post-LGM deglaciation throughout Europe, and by Hughes and Woodward (2017) devoted to the Quaternary glaciations in the Mediterranean Mountains, give excellent overviews on the subject. We prefer to cite few examples here and readers are referred to these comprehensive volumes for further details.

Among these studies the Iberian Peninsula is one of the most studied areas where maximum ice extent was experienced between 35 and 29 ka ago (Serrano et al., 2015). The glaciers advanced again between 23 and 19 ka and ended 18 ka ago. The studies conducted in the Sierra de Gredos Mountains also show that the glaciers reached their maximum extent 26–24 ka ago, began to retreat 21 ka ago with a retreat that accelerated 16 ka ago (Palacios et al., 2011). Similarly, age data from the Sierra de Guadarrama Mountains show that the glacial advance occurred 25–19 ka years ago and the retreat occurred 19–16 ka ago (Palacios et al., 2012). In the Sierra Nevada Mountains the glaciation reached its maximum extent 32–30 ka ago. The glaciers advanced again 20-19 ka ago and the glaciation ended 15–14 ka ago (Gomez-Ortiz et al., 2015). On the other hand on the Cerdanya Massif, the LGM was experienced 23 ka ago with a glacial advance 18–17 ka ago (Palacios et al., 2015a).

The studies conducted on the Alps show that the LGM occurred between 30 and 19 ka. The glaciers that began to retreat in 19–18 ka advanced again around 17–16 ka ago (Ivy-Ochs, 2015). On the Apennines in Italy, the LGM occurred 28–27 ka ago, the retreat began 22–21 ka ago and the advance reoccurred 18 ka ago (Giraudi, 2015).

The studies in the Tatra Mountains in Central Europe revealed that the LGM occurred between 25 and 20 ka. The temperatures were 9–10 °C lower than today with 30–50% less precipitation (Makos, 2015). Deglaciation in the Polski/RoztokiValley in the Tatra Mountains began 18 ka ago and advanced again 12 ka ago (Makos et al., 2013). The glacial advances in the Pietrele Valley in the Southern Carpathians occurred before the LGM, the second advance occurred 16 ka ago and the last advance occurred in the Younger Dryas (Reuther et al., 2007). Deglaciation in the Parang Mountains in Romania began 13 ka ago (Gheorghiu et al., 2015).

The results of the datings conducted on the Chelmos Mountain in Greece reveal that the glacial advance occurred 40–30 ka ago, the retreat occurred 23–21 ka ago and the pause occurred 13–10 ka ago (Pope et al., 2015). The two-stage glacial advance that occurred 24–18 ka ago was determined in the Rila Mountains in Bulgaria (Kuhlemann et al., 2013).

6.5. Rock glacier development

Rock glaciers are typical of periglacial regions and develop with the accumulation of debris material in permafrost areas. In Turkey, rock glaciers are encountered on the İhtiyar Şahap Mountains, Mercan Mountains, Esence Mountains, Taurus Mountains, Erciyes Volcano and especially on the Eastern Black Sea Mountains (Sarıkaya and Tekeli, 2014). Most of the specified rock glaciers have an active character, are located at the bottom of the valleys and have transverse concentric lobes. They are mostly located between 2800 m a.s.l. and 3000 m a.s.l. elevations. However, quantitative age data from these rock glaciers are not yet available with a recent exception of a ^{36}Cl surface exposure dated rock glacier in Mount Geyikdağ in the Taurus Mountain Range (Çiner et al., 2017).

Rock glacier surface exposure dating is in its infancy compared to moraines. Boulders containing major inheritance (Çiner et al., 2017) that yield older ages or problems related to toppling and/or erosion of boulders that result in too young ages (Moran et al., 2016) are reported. The Karçal fossil rock glacier weighted average age of 15.7 ± 1.3 ka is important, as it represents its development during the Late-glacial. However, this age should be considered as a maximum age, as inheritance can be a major problem in rock glacier boulders (see Çiner et al., 2017 for discussion).

The Late-glacial ages obtained from the Karçal fossil rock glacier are compatible with ages determined from some European rock glaciers. For instance, Palacios et al. (2016) reported rock glacier boulder ages of 10.5 ± 0.4 ka, 14.0 ± 0.4 ka and 14.3 ± 0.7 ka from the Aranser Valley in the Southern Pyrenees. In another work carried out on fossil rock glaciers in the Karwendel Mountains on the Alps, Moran et al. (2016) obtained ages from Younger Dryas to the onset of Holocene (12.3 ± 0.6 ka and 10.1 ± 0.6 ka). In a different work reported rock glacier boulder ages change between 14.8 ± 1.9 ka and 11.1 ± 1.3 ka in the upper Gallego Valley (Palacios et al., 2015b). Finally, rock glacier boulder ages are 12.0 ± 0.5 ka, 9.6 ± 0.4 ka, 9.1 ± 0.7 ka and 7.5 ± 0.4 ka in the Sierra Nevada in southern Spain (Gomez-Ortiz et al., 2012). These ages might indicate a preferential timing of rock glacier developments (from Late-glacial to the onset of Holocene), but as previously stated, they should be regarded as maximum ages and preferentially backed up by independent age data.

As a concluding remark it is important to note that the local LGM age (~20 ka) obtained from the recessional moraines in Karçal Valley perfectly fits the regional trend and contributes to the understanding of the glacial geochronology of northeastern Turkey. On the other hand, further studies are needed to confirm whether or not Late-glacial (~15 ka) is a preferential time for rock glacier developments in this part of the world.

7. Conclusion

We presented the first cosmogenic surface exposure ages from the Lesser Caucasus on the Karçal Mountains (41.24° N, 42.06° E, 3431 m a.s.l.), located in the most northwestern part Anatolia. The Karçal Valley hosts a small glacier (2926 m a.s.l.), recent and fossil rock glaciers and lateral and recessional moraines.

Although the maximum extent and timing of the glaciation is not exactly known, as lateral and terminal moraines were not suitable for sampling, recessional moraines indicate that the Karçal Valley palaeoglacier deglaciation started at least 19.9 ± 1.2 ka ago. This age provided a minimum local LGM age of the valley, which is synchronous with the global LGM.

On the other hand, fossil rock glacier boulders were deposited during the Late-glacial (15.7 ± 1.3 ka ago) but rock glacier ages need to be checked by independent age data as they may contain inherited nuclide concentrations. The dates from the Karçal Valley are compatible with other valleys in the nearby Eastern Black Sea and Turkey in general.

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