

Late-glacial and Holocene glacial history of the northeast Mediterranean mountain glaciers - New insights from *in situ*-produced ^{36}Cl -based cosmic ray exposure dating of paleo-glaciers deposits on Mount Olympus, Greece

Michael N. Styllas¹, Irene Schimmelpfennig², Lucilla Benedetti², Mathieu Ghilardi² and ASTER Team^{2*}

¹ GEOSERVICE LTD, Thessaloniki, Greece

² Aix Marseille Univ, CNRS, IRD, INRA, Coll France, CEREGE, Aix-en-Provence, France

* Georges Aumaître, Didier Boulrès, Karim Keddadouche.

Abstract

In this study, we present a new glacial chronology based on 20 *in situ*-produced ^{36}Cl -based cosmic ray exposure datings from moraine boulders and bedrock from the Throne of Zeus (TZ) and Megala Kazania (MK) cirques on Mount Olympus. The ^{36}Cl derived ages of glacial landforms range from 15.6 ± 2.0 to 0.64 ± 0.08 ka, spanning the Late-glacial and the Holocene. The Late-glacial, recorded in both cirques, is partitioned in three distinct phases (LG1-3): an initial phase of moraine stabilization at 15.5 ± 2.0 ka with subsequent deglaciation starting at ~ 14 ka (LG1), followed by a shift to marginal conditions for glaciation at 13.5 ± 2.0 ka (LG2), sustained by large amounts of wind-blown snow, despite regional warming. Glacial conditions returned at 12.5 ± 1.5 ka (LG3) and were characterized by low air temperatures and glacier shrinking. The Holocene glacial phases (HOL1-3) are recorded only in the MK cirque, likely due to its topographic attributes. An early Holocene glacier stillstand (HOL1) at 9.6 ± 1.1 ka follows the regional temperatures recovery. No glacier activity is observed during the mid-Holocene. The Late Holocene glacier expansions, include a moraine stabilization phase (HOL2) at 2.5 ± 0.3 ka, during wet conditions and solar insolation minima, while (HOL3) corresponds to the early part of the Little Ice Age (0.64 ± 0.08 ka). Our glacial chronology is coherent with glacial chronologies from several cirques along the northeast Mediterranean mountains and in pace with numerous proxies from terrestrial and marine systems from the north Aegean Sea.

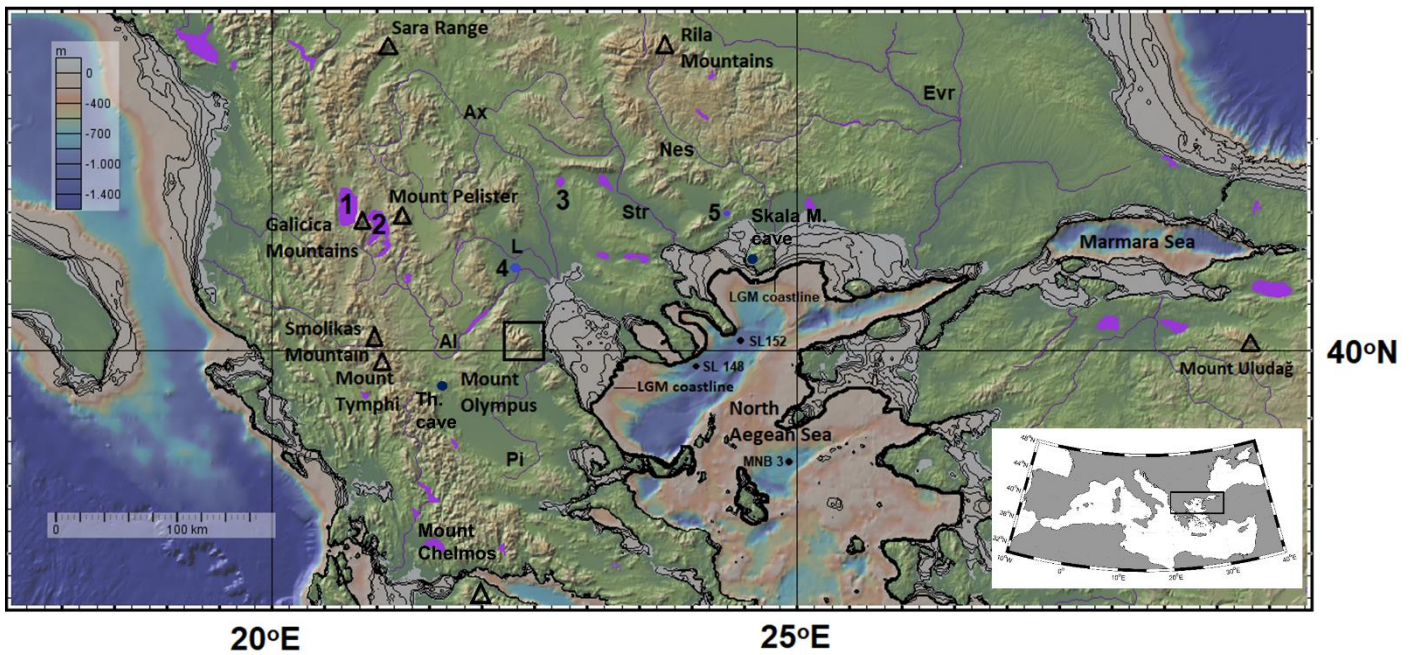
Keywords

Small cirque glaciers;
In situ produces cosmogenic ^{36}Cl ;
Surface Exposure Dating (SED);
Glacial phases;
Late-glacial;
Holocene;
Northeast Mediterranean;
Mount Olympus;
Greece.

57 Glaciers are very good indicators of climate change due to their mass balance sensitivity to variations in
58 precipitation, temperature and solar insolation (Oerlemans, 2005). Therefore, the knowledge of the history
59 of mountain glaciers, as recorded in a variety of glacial landforms, allows for reconstruction of local and
60 regional paleoclimatic conditions. Glaciers usually form above the local climate-dependent equilibrium line
61 altitude (ELA). In climates that are marginal for glaciation, like the ones found in Mediterranean mountains,
62 small cirque glaciers are often formed in protected locations in response to local topoclimatic factors such as
63 excess snow accumulation through the deposition of wind-blown and avalanching snow (e.g. Hughes et al.,
64 2006). Under these conditions, small cirque glaciers can even occur at elevations lower than the ELA, when
65 the accumulation rates are four times higher than the local precipitation (e.g. González Trueba et al., 2008;
66 Hughes, 2009, Huss and Fischer, 2016 and references therein). The advance, stabilization and retreat phases
67 of these small cirque glaciers can thus be triggered by local topoclimatic factors as well as by variations in
68 the regional climate, making them overall invaluable, albeit discontinuous, recorders of past climate
69 variability.

70 The Mediterranean basin with its mid-latitude position and its proximity to the North Atlantic, Eurasian
71 and North African climatic regimes, has undergone significant changes in the terrestrial and marine systems
72 during the late Pleistocene, and this is also reflected in the glacial records of the Mediterranean mountains
73 (e.g. Kuhlemann et al., 2008, Hughes and Woodward, 2008, 2016). The glacial history of the mountains
74 surrounding the Mediterranean basin has been conceived as one of the best recorders of ocean-continent
75 climate interactions and their external forcing mechanisms during periods of major glacier advances as the
76 Last Glacial Maximum – LGM (e.g. Kuhlemann et al., 2008; Domínguez-Villar et al., 2013). To a smaller
77 geographical extent, the sub-region of the northeast Mediterranean (southern Balkans, north Aegean and
78 Marmara Seas, Fig. 1), has been also characterized by complex Late Pleistocene and Holocene marine and
79 terrestrial environmental dynamics, as it comprises a transition region where W-E and N-S contrasting
80 climatic and hydrological regimes collide and interact with each other (e.g. Lawson et al., 2005, Digerfeldt
81 et al., 2007, Kothoff et al., 2008, Marino et al., 2009, Pross et al., 2009, Tzedakis et al., 2009, Schmiedl et
82 al., 2010, Francke et al., 2013, Zhang et al., 2014, Styllas and Ghilardi, 2017, Koutsodendris et al., 2017).

83
84 The recent advances in surface exposure dating (SED) techniques of glacial landforms using *in situ*-
85 produced cosmogenic nuclides such as beryllium-10 (^{10}Be) and chlorine-36 (^{36}Cl) (e.g. Balco, 2011), have
86 resulted in an increasing number of studies that have considerably improved our understanding of the Late
87 Pleistocene glacial extents and timing of Mediterranean mountains (Hughes and Woodward, 2017 and
88 references there in). In the late 1990's Mount Olympus, Greece's highest mountain, saw one of the earliest
89 attempts for dating glacial deposits along the Mediterranean mountains using ^{36}Cl (Manz, 1998). The results
90 of more recent studies that followed this pioneering attempt, suggest that the Mediterranean paleoglaciers
91 generally advanced during two periods after the global LGM (*c.* 27.5–23.3ka, Hughes and Gibbard, 2015),
92 confined between 16-15 ka and 13-10 ka, and in phase with the GS-2a and GS-1 stadials in the Greenland
93 oxygen isotope record (Ribollini et al., 2017 and references therein).



95
 96 **Fig. 1.** Location of the study area (black rectangle) and of the northeast Mediterranean mountains (black triangles) where
 97 cosmogenic surface exposure dating of glacial landforms (SED), or glacier equilibrium line altitude (ELA) reconstructions are
 98 available since the Last Glacial Maximum (LGM). Purple areas correspond to major lakes, while more specifically, numbers 1, 2,
 99 3, 4 and 5 correspond to Lakes Ohrid, Prespa, Dorjan, Loudias and Tenaghi Phillipon swamp. Also shown are the main fluvial
 100 systems discharging into the north Aegean Sea (Pi: Pinios River, Al: Aliakmon River, L: Loudias River, Ax: Axios River, Str:
 101 Strymonas River, Nes: Nestos and Evros/Meriç River). Blue circles show the locations of speleothem and sediment cave
 102 records (Th: Theopetra Cave, Duhlata Cave, Skala M: Skala Marion Cave). The locations of north Aegean marine cores SL 152,
 103 SL 148 and MNB 3 are also shown (black dots). The bathymetric contours between -120 and -20m, representing the LGM (thick
 104 black line) and early Holocene coastlines, are also shown. Topographic and bathymetric background is provided by Geomapapp@
 105 (<http://www.geomapapp.org>, Ryan et al., 2009).
 106

107 Accordingly, during the last decade, an augmented volume of research based on geomorphological
 108 evidence, paleoclimatic reconstructions and SED of glacial deposits and landforms, allowed assessing a
 109 general framework of the glacial evolution of the southern Balkan and northwestern Turkey mountains
 110 during the LGM and the Late Glacial (e.g. Pindos Range: Tymphi Mountain and Mount Smolikias – Hughes
 111 et al., 2006, Sara Range – Kuhlemann et al., 2009, Mount Uludağ - Zahno et al., 2010, Rila Mountains -
 112 Kuhlemann et al., 2013, Mount Chelmos – Pope et al., 2015, Mount Pelister – Ribollini et al., 2017, Galicica
 113 Mountains – Gromig et al., 2017, Fig. 1). With the exception of Mount Uludağ in NW Anatolia, the number
 114 of SED studies in the northeast Mediterranean mountains is, however, still limited and does not provide a
 115 well-constrained glacial chronology.

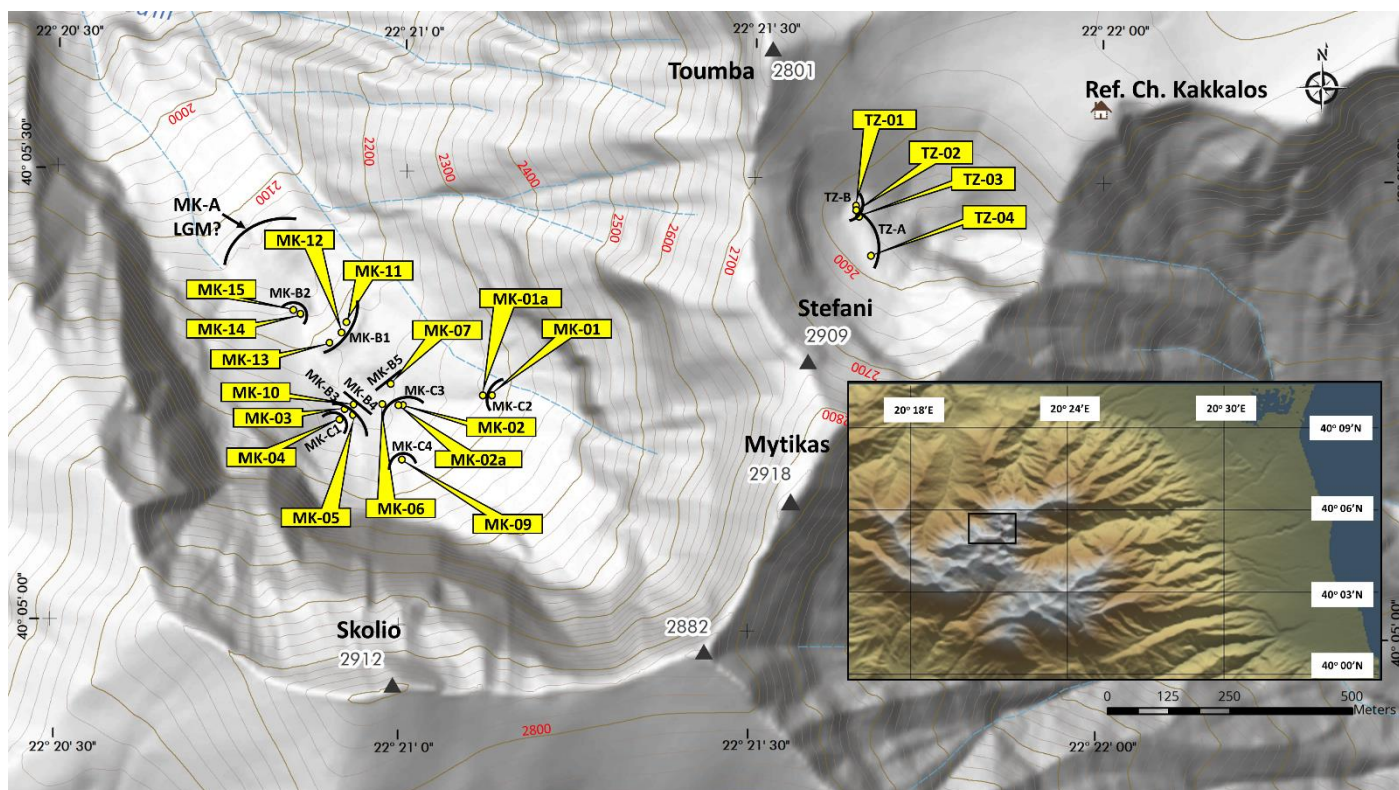
116 Despite the fact that glaciers are uncommon in Greece, we present here for the first time a glacier
 117 chronology spanning the Late-glacial and the Holocene, derived from ³⁶Cl in situ cosmogenic dating of two
 118 small (< 0.5km²) cirque glaciers on Mount Olympus, based on 20 rock samples from bedrock and glacially
 119 transported boulders (Fig. 2). Our glacial chronology is compared to the existing SED studies from glacial
 120 cirques situated along the headwaters of the northeast Mediterranean Sea (north Aegean and Marmara seas),
 121 in an effort to reconstruct a regional and robust signal of glacier fluctuations. By correlating our findings
 122 with well-studied terrestrial (lacustrine, fluvial sequences and speleothems) and marine records from the
 123 same region (Fig. 1), we interpret our new glacier chronology in terms of external and the underlying local
 124 and regional climate forcing. The density of sampled boulders in Megala Kazania cirque further allows us to
 125 propose a chronology of glacier oscillations, to depict paleoclimatic information and to correlate the findings
 126 with glacier and other proxy records from other northeast Mediterranean.
 127

128 **2. Climatic and glacio-geomorphic setting of Mount Olympus**

129 *2.1 Past and present climatic conditions*

130 Mount Olympus is a coastal massif of circular shape, composed of Triassic and Cretaceous carbonate
 131 sequences, rising 2918m above the northwestern Aegean Sea coastline (Fig. 1, Fig. 2 inset map). Its

132 proximity to the sea as expressed by the short distance (18km) of its highest peaks from the shore, has a
133 pronounced impact on the local climate, with increased supply of moisture and high precipitation and
134 temperature gradients (Styllas et al., 2016).
135



136 **Fig. 2.** Locations of the respective groups of moraines (MK-A, MK-B1-5, MK-C1-4, TZ-A and TZ-B), illustrated with black
137 curved lines and of the boulder and bedrock samples ($n=20$), illustrated in yellow dots and boxes selected for *in situ*-produced
138 ^{36}Cl -based cosmic ray exposure dating from the Throne of Zeus (TZ) and Megala Kazania (MK) cirques. (DEM source:
139 Ktimatologio A.E. Elevation contours provided from RouteMaps.gr, Inset map data: SRTM 90).
140
141

142 The high plateaus and valleys of Mount Olympus exhibit a variety of glacial and periglacial landforms and
143 deposits, testifying to the recurrent presence of Pleistocene glaciations, which tentatively have been ascribed
144 to the marine isotope stages (MIS) 8, 6 and 4-2 (Smith et al., 1997). During the first stage of the last
145 glaciation (Würmian MIS 4-2), the snowline was depressed down to 1300m, and during the latter stage of
146 MIS 2 (LGM) it was confined within the upland cirques at elevations higher than 2000m (Smith et al., 1997,
147 2006). Within one of these cirques, the Megala Kazania cirque, extant permanent snowfields and ice bodies
148 under the scree slopes have been considered remnants of the final Late Holocene deglaciation, which most
149 likely occurred after the Little Ice Age (Styllas et al., 2016).

150 During the LGM, the overall conditions in the vicinity of Mount Olympus were substantially different, as
151 the sea-level of the Aegean Sea was 120m lower (Lambeck 1996), and large portions of the present day
152 continental shelves were exposed subaerially (Perissoratis and Conispoliatis, 2003), reducing significantly
153 the overall water surface area of north Aegean Sea (Fig. 1). The LGM Aegean Sea Surface Temperatures
154 (SST's) were 6°C lower compared to present values (Kuhlemann et al., 2008), and the local ELA was at
155 $2000 \pm 150\text{m}$; a depression that resulted from a significant drop ($\sim 7\text{--}9^{\circ}\text{C}$) in air temperatures (Peyron et al.,
156 1998, Kuhlemann et al. 2008). In a study by Styllas et al., (2016), geomorphological evidence and Holocene
157 paleoclimatic simulations for Megala Kazania cirque, based on projected values of annual precipitation
158 (P_{ann}) and summer temperature (T_s) from a proximal (13km to the north) meteorological station, suggest that
159 the best candidates for a glacier to exist and/or to advance with the local ELA¹ at 2200m, are $P_{ann} = 2160$
160 $\pm 160\text{mm}$ and $T_s = 4.6 \pm 0.4^{\circ}\text{C}$. These values are very similar to the ones found ($P_{ann} = 2300 \pm 200\text{mm}$ and T_s
161 $= 4.9^{\circ}\text{C}$) for the local glacial maximum of Pindos Mountains (Mount Tymphi, Fig. 1) during the Tymphian
162 (28.2 – 24.3 ka BP) cold stage (Hughes et al., 2003, 2006a, 2008). Therefore, one would expect that the
163 considerable drop in air temperatures ($\sim 7^{\circ}\text{C}$) during the LGM would have resulted in a significant drop in
164 the local ELA, in both Mount Olympus and Mount Tymphi, but the reconstructed drop in the local ELA

¹ Given the small vertical range ($\sim 200\text{m}$) and size ($< 0.5\text{km}^2$) of TZ and MK glaciers, we consider that the observed frontal moraines elevation coincides with the local ELA.

165 derived from a regional synthesis for the entire Mediterranean Basin, was only 200m (Kuhlemann et al.,
166 2008). This can be explained by the fact that during the LGM, both Mount Olympus and Mount Tymphi
167 were out of the zone of major Mediterranean cyclone tracks, which were deflected either further north
168 towards the Adriatic basin and the western Balkans, or further south towards the eastern Mediterranean and
169 resulted in drier conditions over the northeast Mediterranean (Kuhlemann et al., 2013 and references
170 therein). The presence of rock glaciers on Mount Tymphi down to 1800m is consistent with the LGM
171 temperature depressions and moisture deficiency (Hughes et al., 2003). Superimposed on the LGM
172 atmospheric circulation patterns of low cyclonic activity over the southern Balkans and the Aegean Sea, was
173 the fact that during the LGM Mount Olympus was distanced 100km from the coastline (Fig. 1). Potentially,
174 the distant cold waters and the overall reduced water surface area of the north Aegean Sea, also resulted in
175 moisture deficit and reduced precipitation, conditions more representative of a continental setting that was
176 restrictive to large glacier expansions. Therefore, the altitudinal extent of our study sites, is confined above
177 the reconstructed LGM ELA ($2000 \pm 150\text{m}$, moraine MK-A, Fig. 2) and is considered to correspond to the
178 range of Late-glacial and Holocene ELAs.

179 At present, the distribution of precipitation in the vicinity of Mount Olympus varies with altitude and
180 distance from the coast, as the highest peaks constitute an orographic barrier, which results in a climatic
181 partition between the eastern (marine) and western (continental) sides of the mountain (Styllas et al., 2016).
182 Along the eastern and western piedmonts and at elevations between 50-150m, annual precipitation (P_{ann})
183 and temperature (T_{ann}) are 550mm and 15°C respectively, whereas closer to the mountain and at elevations
184 between 400 and 800m, P_{ann} reaches 600–800mm and T_{ann} is 9.9°C (Styllas et al., 2016). The eastern
185 (marine) side of Mount Olympus receives on average 200mm more precipitation than the western
186 (continental) side (period of observations 1960 – 2000, Styllas et al., 2016). Above 1500m, most of the
187 winter (December - March) precipitation occurs as snow, and in depressions below the highest peaks,
188 including the cirques under consideration, the snowpack thickness reaches excess values of 1.5m. Mean
189 summer (June – September) temperatures recorded in the meteorological station of refuge Christos Kakkalos
190 (Fig., 2, period of observations 2007 – 2016), range between 8.5 and 10.5°C . Winter precipitation (P_w) on
191 Mount Olympus is mainly related to cyclogenesis over the Aegean Sea and central Mediterranean (e.g.
192 Flockas et al., 1996, Batzokas et al., 2003), while on interdecadal timescales the North Atlantic climate
193 exerts strong control on P_w , as indicated by the high negative correlations between P_w with the winter
194 (December – March) North Atlantic Oscillation (wNAO) index (Styllas et al., 2016).

195 2.2 Glacial geomorphology

196 One of the most prominent features of Mount Olympus topography known to humans since antiquity, is the
197 east facing Throne of Zeus cirque (TZ), escarped under the 150m high rock cliff of Stefani (meaning the
198 crown in *Greek*, Fig. 3). Two well-preserved moraines (TZ-A, TZ-B), situated nearly perpendicular to each
199 other, rest at an elevation of 2580m, bounding a surface area of 0.150 km^2 between the rock cliff base and
200 the moraine crests (Fig. 3). TZ-A is a 2-4m high frontal moraine with an N-S orientation, parallel to the cliff
201 base, located above a rock step. Below moraine TZ-A, the 30m high exposed rock step is indicative of a
202 previous, likely a LGM glacier retreat, based on direct observations of glacial striations. Moraine TZ-A is
203 morphostatigraphically overlain by a series of periglacial deposits and also by moraine TZ-B, an E-W
204 oriented moraine with a 5-8m high well-developed crest. The periglacial deposits overlying moraine TZ-A
205 include protalus ramparts, talus slope deposits and rockfall talus deposits, while below the characteristic
206 rock band that separates the upper scree field with the underlying solifluction lobes, fresh talus deposited in
207 the form of debris fans, are indicative of the high rates of the cliff discharging ability and sediment
208 mobilization (Fig. 3). These periglacial deposits overlying moraine TZ-A in a stratigraphic upward sequence
209 (Fig. 3 and 6A), suggest that the high supply rates of rock material, were likely forced by a change of the
210 climate conditions from glacial to periglacial after the stabilization of TZ-A, however the timing remains
211 unknown. In this east facing setting, direct inputs of solar insolation and high temperature differences
212 between day and night are considered the main driving mechanisms for the observed high rates of physical
213 weathering and the resultant high amounts of debris. The intense frost shattering leads to a continuous
214 renewal of the cliff surface, thus the possibility of dating boulders with significant prior exposure from the
215 cliff above is extremely low (Putkonen and Swanson, 2003).

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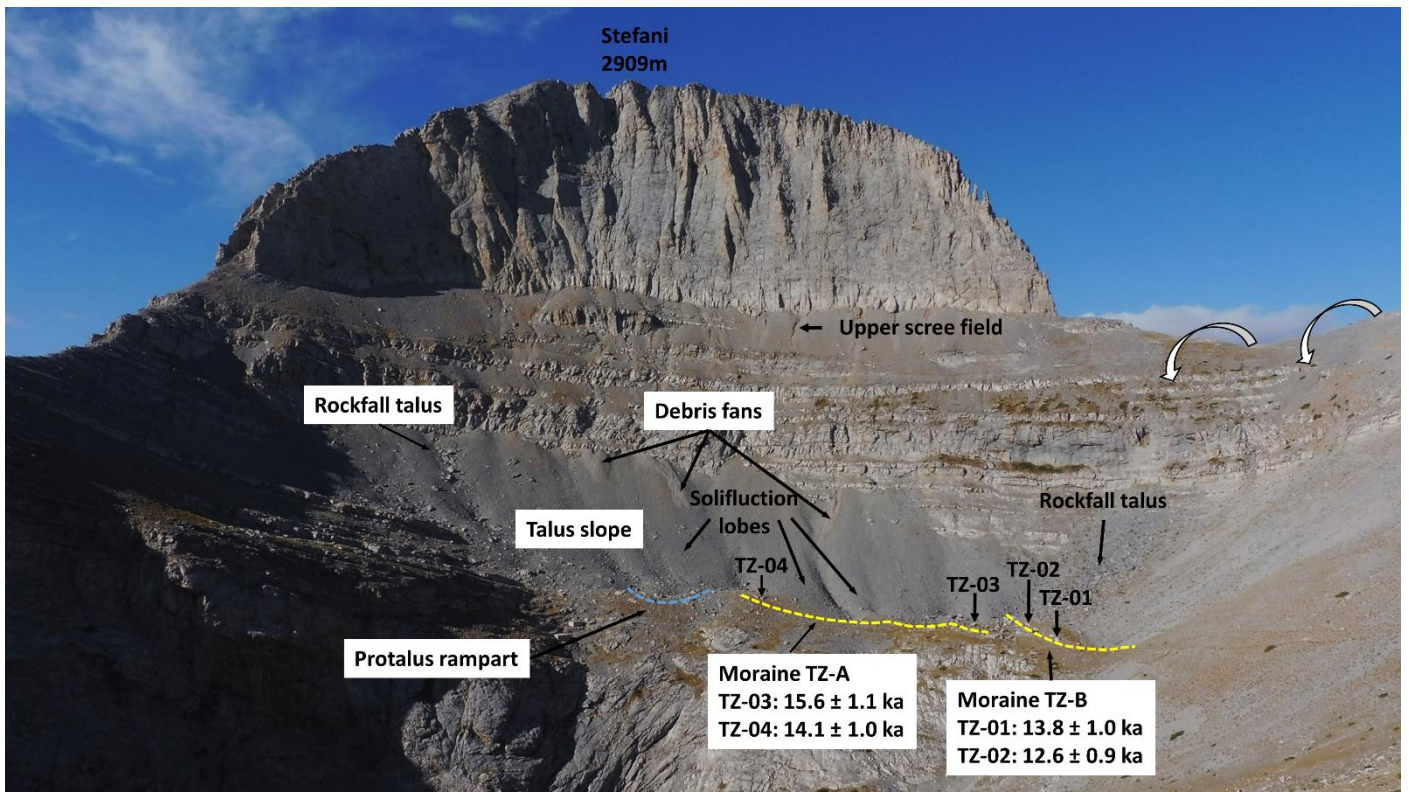


Fig. 3. The east facing Throne of Zeus (TZ) cirque with the 150m high headwall of Stefani (2909m). Moraines TZ-A and TZ-B are highlighted in yellow color. The locations and ages of the sampled boulders (#TZ-01–03) and bedrock (# TZ-04) are shown, with analytical errors only, and with the main periglacial features. The white curved arrows denote the direction of windblown and avalanching snow entering the cirque from the saddle above moraine TZ-B under W, SW and/or NW winds (direct observations).

The largest glacial landform of Mount Olympus and of Greece in general, is the NW-oriented head-valley cirque of Megala Kazania (MK), bounded by Mount Olympus highest peaks (Fig. 4). The cirque floor hosts three well-preserved groups of moraines (MK-A, MK-B and MK-C). Moraine group MK-A being located farthest from the cirque headwall, i.e. it is the oldest, and MK-C is the nearest, thus youngest moraine complex in the sequence (Styllas et al., 2016; Figs 2, 5). The cirque area between frontal moraine MK-A and the cirque headwall, is 0.5 km². MK-A (2155m) is overlain by an undated soil horizon (Sa) and fan-type glaciofluvial (Gf) deposits (Fig. 4). MK-B is bounded by the most characteristic and well-preserved moraine found on Mount Olympus (MK-B1, crest elevation 2225m; Fig. 5), a 65m-high frontal moraine with a steep distal slope (Fig. 4), inbound of which several other soil horizons are observed. Upslope of MK-B1, a variety of glacial landforms (multiple moraine crests, sinkholes, block accumulations, polished rock and striated surfaces) point to multiple episodes of glaciation (Figs 4 and 5), while in the downslope direction, MK-B1 is dissected by MK-B2 (Figs 4, 5, 8), which we tentatively interpreted as a push moraine; a landform interpreted by Smith et al. (1997) as a protalus rampart, during their field reconnaissance. Such talus-derived landforms have been associated with periglacial conditions and the formation of rock glaciers (Shakesby et al., 1987). However, the distinction between relict pronival-protalus ramparts and push moraines in the Alpine realm is not clear and the morphogenetic boundaries between the two often overlap (Scapozza, 2015), but here, we consider MK-B2 a push moraine, as we did not find additional evidence for the existence of an active rock glacier.

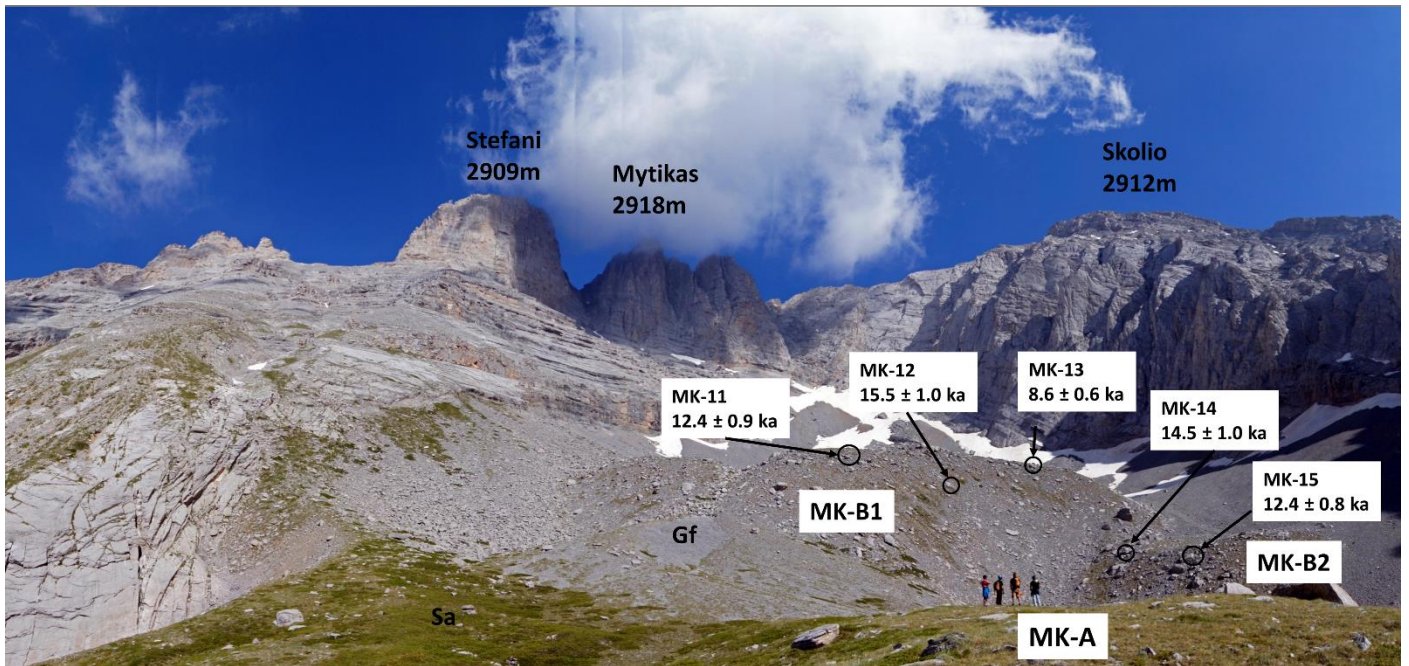


Fig. 4. Panoramic view of Megala Kazania (MK) cirque, with the most characteristic moraine MK-B1 in the foreground and Mount Olympus highest peaks and perennial snowfields in the background. The photograph is taken from morainic complex MK-A (white dot in Fig. 5), with the people standing on its crest. MK-A is stratigraphically overlain by the soil horizon Sa (with the grass cover) and by the 65m high moraine MK-B1, breached along its lateral sides by gravely glaciofluvial deposits (Gf) and by the push moraine MK-B2. The locations and ages of the sampled boulders # MK-11–15 are also shown, with analytical errors only.

Stratigraphically inboard of moraine MK-B1, two main glacial landforms are present: an apron of angular and sub-angular large block deposits (block sizes >0.5m), with a high degree of fragmentation and a series of transversal and longitudinal short and 5-7m high crests, sinkholes and near conical mounds of debris (Fig. 5). The transversal crest MK-B4 (2235m) and longitudinal crest MK-B5 (2245m), both of minimal curvature (Fig. 5, 8, 9A and 9B) and located closer to the cirque center than moraine MK-B3, are interpreted as a set of hummocky moraines. Behind these landforms and closer to the cirque walls, a series of blocky moraines, ascribed to morainic complex MK-C. MK-C group of moraines are 2 – 4m high and less developed compared to MK-B moraine crests (Fig. 5), suggesting shorter periods of glacier-friendly climatic conditions and moraine stabilization. Blocky moraines MK-C1 – 2240m, MK-C2 – 2290m, MK-C3 – 2250m and MK-C4 – 2270m, are either nested within older MK-B moraines (as for example moraine MK-C1 is nested in moraine MK-B3 – 2240m; Fig. 5), or breached by glaciofluvial activity (MK-C2), or located higher up and very close to the cirque walls (MK-C4), where extant snowfields still survive (Fig. 5). Inboard (upvalley) of moraine MK-C3, isolated boulders deposited along a longitudinal ridge parallel to the cirque centerline, are interpreted as supraglacial deposition during the latter stages of MK glacial activity. In contrast to the color of the MK-B deposits, the boulders of moraine complex MK-C are characterized by darker color and show no degree of fragmentation (Fig. 5).

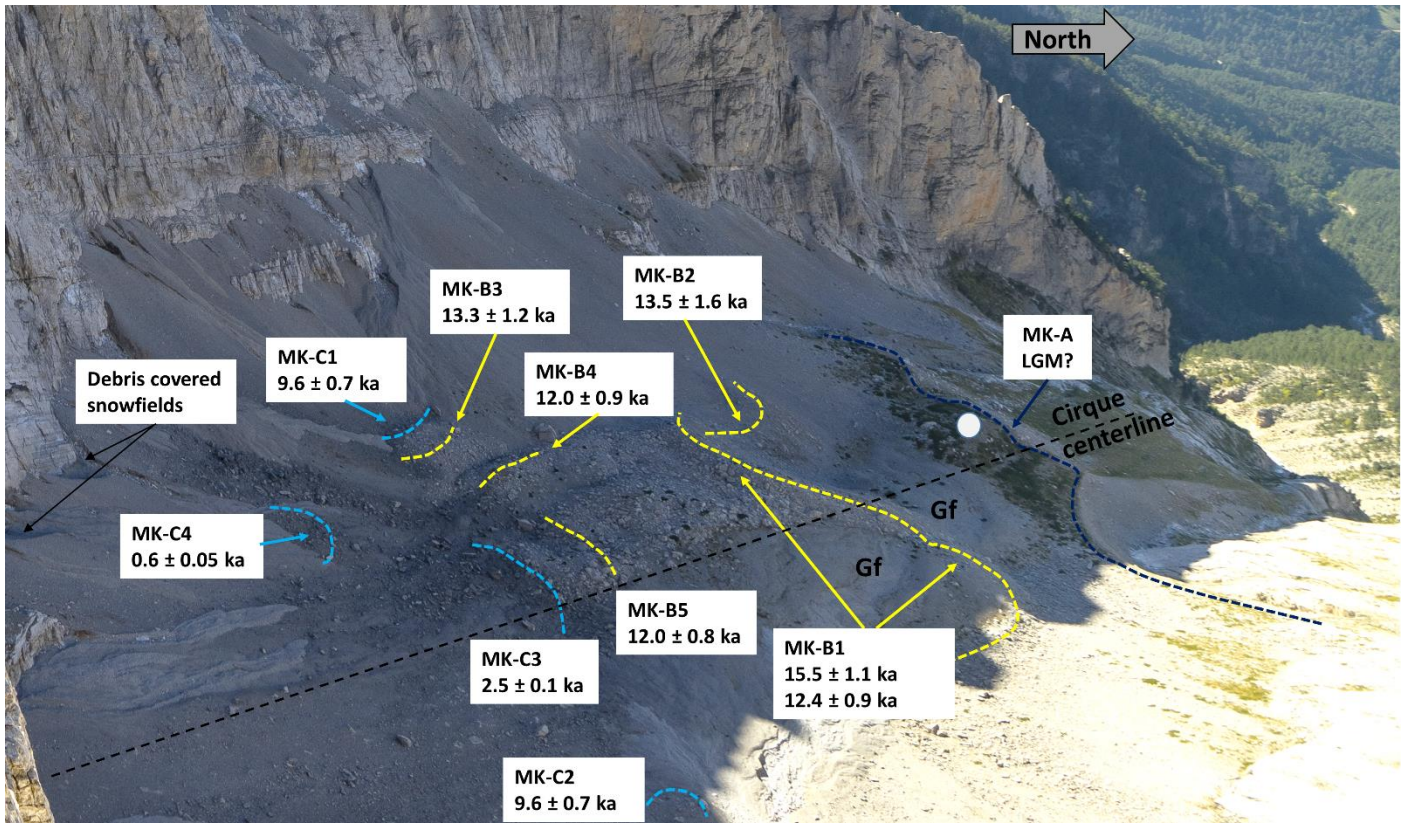


Fig. 5. The view of Megala Kazania (MK) cirque from the summit of Mount Olympus, Mytikas (2918m). The photograph was taken on October 1st 2016, at the beginning of the snow accumulation season, when debris-covered extant snowfields still survive in the shaded parts of the cirque below the steep cliffs of Skolio (2912m). Moraine complexes MK-A (dark blue dashed line), MK-B (yellow dashed lines) and MK-C (light blue dashed lines) and sampled moraine crests (MK-B1–B5, MK-C1–C4) are shown with their respective ³⁶Cl ages, either based on a single sample, or on the arithmetic mean of two or three samples were available. For intersite comparisons between the various landforms of TZ and MK cirques, we do not include the production rate errors in the age uncertainties shown here. Outwash plain glaciofluvial deposits (Gf) that breach the (north) right side of moraine MK-B1 along the cirque centerline, and the boulder apron bounded between hummocky moraines MK-B4, MK-B5 and frontal moraine MK-B1, are also illustrated. The white dot represents the location, where the people are standing in Fig. 4.

3. *In situ*-produced ³⁶Cl-based cosmic ray exposure dating of Throne of Zeus (TZ) and Megala Kazania (MK) moraines

3.1 Fieldwork and collection of samples

During summer and fall of 2015 and 2016, a fieldwork reconnaissance took place and initially involved the detailed mapping of all glacial features (moraines, striated bedrock and glacial sediments) in both cirques. In addition to the mapping of the moraines, we collected 20 samples (16 samples from MK and 4 samples from TZ, Fig. 2, and Table 1). One sample was retrieved from bedrock (sample TZ-04) and the other 19 samples from glacially transported boulders.

Table 1

Geographical characteristics, boulder heights and topographic shielding and snow cover correction factors for the ³⁶Cl-dated samples

Sample & Landform ID	Latitude °N (DD) WGS 84	Longitude °E (DD) WGS 84	Landform Elevation (m)	Boulder height (m)	Shielding Factor	Snow correction factor	Boulder location
Moraine TZ-A			2592			0.94	
TZ-03	40.0910	22.3609	2597	0.75	0.931		Moraine crest
TZ-04	40.0903	22.3612	2585	1.25	0.931		Bedrock inboard of moraine

Moraine TZ-B			2580			0.94	
TZ-01	40.0912	23.3608	2587	1.0	0.932		Moraine crest
TZ-02	40.0911	23.3608	2575	0.6	0.946		Moraine crest
Moraine MK-B1			2225			0.94	
MK-11	40.0889	22.3487	2220	0.6	0.894		Embedded on outer slope below moraine crest
MK-12	40.0885	22.3483	2215	0.5	0.902		Embedded on outer slope below moraine crest
MK-13	40.0887	22.3486	2223	0.7	0.906		Deposited on outer slope below moraine crest
Moraine MK-B2			2175			0.94	
MK-14	40.0890	22.3476	2170	1.2	0.871		Embedded in protalus rampart
MK-15	40.0891	22.3474	2167	0.9	0.884		Embedded in protalus rampart
Moraine MK-B3			2240			0.90	
MK-03	40.0872	22.3489	2235	0.6	0.820		Embedded in moraine crest
MK-05	40.0873	22.3487	2240	0.7	0.844		Top of moraine crest
Moraine MK-B4			2245			0.94	
MK-10	40.0874	22.3489	2245	0.9	0.852		Top of hummocky moraine
Moraine MK-B5			2245				
MK-07	40.0878	22.3498	2245	0.4	0.881		Top of hummocky moraine
Moraine MK-C1			2240			0.94	
MK-02	40.0874	22.3501	2238	2.0	0.864		Inbound of moraine crest
MK-02a	40.0874	22.3500	2236	1.5	0.864		Inbound of moraine crest
MK-06	40.0874	22.3496	2236	0.3	0.852		Embedded in outer slope of moraine
Moraine MK-C2			2290			0.90	
MK-04	40.0871	22.3486	2290	0.9	0.821		Crest top of moraine MK-C2, nested in MK-B3
Moraine MK-C3			2250			0.90	
MK-09	40.0864	22.3501	2250	1.1	0.818		On top of blocky moraine
Moraine MK-C4			2300			0.90	
MK-01	40.0876	22.3522	2300	1.8	0.848		On top of breached lateral moraine
MK-01a	40.0876	22.3520	2300	3.0	0.848		On top of breached lateral moraine

* MK: Megala Kazania cirque, TZ: Throne of Zeus cirque

The samples from the TZ and MK cirques were retrieved using a hammer and a chisel from flat or slightly tilted surfaces of selected limestone boulders that were either embedded or rested on the top of eleven moraine crests (TZ-A–B, MK-B1–B5 and MK-C1–C4, Fig. 2, 3 and 4). Their coordinates were logged in a hand-held GPS (Garmin E-trex). Boulders with evidence of post depositional movement (toppling) were avoided as were boulder surfaces with high degree of fragmentation and signs of surface and rill erosion. The boulders' heights, landform elevations and the surrounding topographic shieldings are given in Table 1.

3.2 Sample preparation and exposure ages calculations

The samples were physically and chemically prepared at the Centre Européen de Recherche et d'Enseignement des Geosciences de l'Environnement (CEREGE) in Aix en Provence, France, following

standard procedures (modified from Stone et al., 1996; Schlagenhauf et al., 2010). After mechanical crushing and sieving of the 20 samples, ~35 g of the 0.25–0.5mm grain size fractions were first washed in H₂O and then leached in 2M HNO₃ in order to remove atmospheric and potential secondary Cl. After total dissolution of the leached grains in 2M HNO₃, a ³⁵Cl-enriched spike solution containing ~1.9 mg Cl was added to the sample solutions. From this step on, two chemistry blanks were processed, one for each batch of 10 samples. The solutions were filtered to remove any solid residues. From these solutions, an 1 ml aliquot was taken to later conduct measurements of the Ca concentrations in the dissolved samples by inductive coupled plasma atomic emission spectrometry (ICP-AES) at CEREGE (Table 2). 2 ml of a 10% AgNO₃ solution were added to the remaining filtered solutions to precipitate AgCl, and after two days the supernatant was removed by pumping and centrifuging. After re-dissolving the AgCl in NH₃aq, the removal of S from the solution was achieved through addition of 0.5 ml of a saturated Ba(NO₃)₂ solution, which resulted in precipitation of BaSO₄. The remaining solutions were filtered with syringes through 0.45µm-mesh filters, and AgCl was re-precipitated by adding ~2 ml of concentrated HNO₃. The rinsed and dried AgCl pellets were pressed into nickel cathodes and measured by isotope dilution accelerator mass spectrometry (AMS) (Bouchez et al. 2015, Ivy-Ochs et al., 2004) at French AMS national facility ASTER at CEREGE (Arnold et al., 2013). Both the ³⁶Cl/³⁵Cl and the ³⁵Cl/³⁷Cl ratios were obtained by normalization to inhouse standard SM-CL-12 with an assigned ³⁶Cl/³⁵Cl value of $(1.428 \pm 0.021) \times 10^{-12}$ (Merchel et al., 2011), and assuming a natural ³⁵Cl/³⁷Cl ratio of 3.127. From these measured ratios, the samples' ³⁶Cl and Cl concentrations were calculated (Table 2).

The ³⁶Cl ages were calculated using the Excel® spreadsheet published in Schimmelpfennig et al. (2009), using the ³⁶Cl production rate for spallation of Ca, referenced to sea level and high latitude (SLHL), of 42.2 ± 4.8 atoms ³⁶Cl (g Ca)⁻¹ yr⁻¹ (Schimmelpfennig et al., 2011) with the time-invariant scaling method by Stone (2000). This SLHL ³⁶Cl production rate for spallation of Ca was calibrated with Ca-rich feldspars from lava surface samples collected at Mount Etna volcano, which is the closest to Mt Olympus among all currently existing ³⁶Cl calibration sites. In addition, its value is supported by the SLHL ³⁶Cl spallation production rate inferred from fitting modelled ³⁶Cl data to ³⁶Cl measurements from a calcium carbonate depth profile in south-eastern France (Braucher et al., 2011). The production rate of epithermal neutrons from fast neutrons in the atmosphere at the land/atmosphere interface of 696 ± 185 neutrons (g air)⁻¹ yr⁻¹ (Marrero et al., 2016) was applied. Due to the low Cl concentrations in these limestones (5-23 ppm), the contribution of ³⁶Cl production from low-energy-neutron capture by ³⁵Cl accounts for only between 1% and 6% of the total ³⁶Cl production. A high-energy neutron attenuation length of 160 g cm⁻² was used. We assumed a bulk rock density of 2.7 g cm⁻³ for all samples. Age uncertainties correspond to 1σ; in Table 3 we show the full uncertainties in the individual ages (both analytical and production rate errors are propagated through) as well as the analytical uncertainties only. For the moraines that contain more than one dated boulder, we tested the consistency of the ages through the χ^2 criterion (Ward and Wilson, 1978, outliers are shown in Table 3) and we then calculated the arithmetic mean to obtain the moraine age (Table 3, Fig. 7). Unless otherwise stated, the uncertainties in the mean moraine ages include the standard deviation and the analytical and production rate errors added by propagation in quadrature.

3.3 Corrections for snow and denudation

The calculated SED ages were corrected for snow cover, which is not temporally uniform among the landforms under consideration, according to our modern yearly observations. The studied moraines were therefore divided into two groups. Group 1 comprises moraines MK-B3, -C1, -C2 and -C4 (Fig. 5), which are situated close to the high MK cirque cliffs and are affected by the pronounced effects of windblown and avalanching snow from the cliffs above; they are snow-covered for ~7 months per year. Group 2 includes moraines TZ-A, -B and moraines MK-B1, -B4, -B5, -C3 and push moraine / protalus rampart MK-B2, which are distant from the MK cirque cliffs and are exposed ~2 months earlier during the snowmelt season than the moraines of group 1. It has to be noted that correcting cosmogenic nuclide production rates for the effect of snow cover is challenging for two reasons. First, the exact quantity and yearly duration of snow cover during the period of exposure of the rock surface is generally not known; and second, it is still debated how to correctly model the effect of snow on the different production reactions of the various cosmogenic nuclides, in particular those of multi-reaction-produced ³⁶Cl (Masarik et al., 2007; Zweck et al., 2013; Dunai et al., 2014; Delunel et al., 2014).

As the ³⁶Cl in our samples is dominantly produced by spallation of Ca, we approximate snow correction factors for both above-mentioned moraine groups in a first step, based on the classical equation for the effect

365 of snow cover by Gosse and Phillips (2001), which assumes that the snow pack shields the rock surface and
366 therefore reduces the spallogenic cosmogenic nuclide production compared to no-snow-conditions. For this
367 calculation, we consider a snow pack of 2 m with a snow density of 0.2 g cm^{-3} for 7 and 5 months for
368 moraine groups 1 and 2, respectively. However, we suspect that the snow correction factors of 0.87 and 0.91
369 resulting from these estimates might overestimate the impact of snow, because the slight presence of ^{35}Cl in
370 our samples (section 3.2) results in the opposite of the shielding effect on the spallogenically produced ^{36}Cl ;
371 i.e. hydrogen in snow enhances the low-energy neutron flux at the air/rock interface and thus increases the
372 ^{36}Cl surface production from low-energy-neutron capture by ^{35}Cl (Masarik et al., 2007; Zweck et al., 2013;
373 Dunai et al., 2013). In a second step, we therefore estimate a related ^{36}Cl production increase of 3%,
374 approximated from the correction factors proposed in Fig. 7 of Dunai et al. (2014), resulting in total snow
375 correction factors of 0.90 for moraine group 1 and 0.94 for moraine group 2, i.e. the ^{36}Cl production is
376 reduced by 10% and 6%, respectively, compared to no-snow-conditions.

377 In the lack of quantitative estimates for denudation of the calcareous rock formations in the southern Balkan
378 region and taking into account the fact that the sampled boulders showed minimal signs of surface and/or rill
379 erosion, we discuss our ^{36}Cl ages without any correction for denudation. These therefore correspond to
380 minimum exposure ages. However, in Table 3 we also provide the ^{36}Cl ages including a denudation rate of
381 5mm/ka, as this value was applied in the calculation of ^{36}Cl moraine ages in a recent study from the Balkan
382 region (Gromig et al., 2017). Anyway, the exposure ages calculated both ways for all samples are not
383 significantly different taking into account the associated uncertainties.

384 **Table 2.** Chemical data and AMS measurement results.
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Sample & Landform ID	Sample weight (g)	ASTER cathode	Mass of Cl in spike (mg)	³⁵ Cl/ ³⁷ Cl	³⁶ Cl/ ³⁵ Cl (10 ⁻¹³)	Cl concentration (ppm)	³⁶ Cl blank correction	³⁶ Cl concentration (10 ⁵ atoms g ⁻¹)	CaO concentrations (%)
Moraine TZ-A									
TZ-03	36.28	BVOY	1.893	17.94 ± 0.24	14.79 ± 0.53	13.85 ± 0.23	0.28%	15.98 ± 0.58	53.87 ± 0.64
TZ-04 (bedrock)	35.74	BVON	1.585	10.596 ± 0.099	13.92 ± 0.48	23.30 ± 0.32	0.23%	14.97 ± 0.52	53.59 ± 0.58
Moraine TZ-B									
TZ-01	35.94	BVOX	1.891	29.8 ± 0.52	14.36 ± 0.50	7.38 ± 0.16	0.31%	14.42 ± 0.51	54.70 ± 0.62
TZ-02	35.05	BVOM	1.833	16.93 ± 0.24	11.99 ± 0.42	14.30 ± 0.27	0.27%	13.15 ± 0.46	52.85 ± 0.53
Moraine MK-B1									
MK-11	33.44	BVOT	1.872	31.17 ± 0.66	10.03 ± 0.40	7.43 ± 0.20	0.45%	10.65 ± 0.43	56.73 ± 0.68
MK-12	35.30	BVOU	1.870	23.11 ± 0.38	9.64 ± 0.33	10.20 ± 0.21	0.45%	10.09 ± 0.35	44.03 ± 0.45
MK-13 (<i>outlier</i>)	34.68	BVOV	1.886	26.44 ± 0.38	6.76 ± 0.27	8.21 ± 0.16	0.50%	7.11 ± 0.29	55.68 ± 0.78
Moraine MK-B2									
MK-14	34.80	BVOW	1.810	24.99 ± 0.36	11.03 ± 0.39	9.07 ± 0.17	0.41%	11.19 ± 0.39	55.77 ± 0.56
MK-15	35.42	BVOJ	1.886	40.88 ± 0.49	9.66 ± 0.33	5.038 ± 0.084	0.47%	9.51 ± 0.33	55.51 ± 0.60
Moraine MK-B3									
MK-03	34.37	BVOQ	1.827	30.07 ± 0.59	8.88 ± 0.36	7.36 ± 0.18	0.52%	8.99 ± 0.36	55.88 ± 0.63
MK-05	35.00	BVOS	1.870	27.02 ± 0.31	10.05 ± 0.40	8.47 ± 0.12	0.44%	10.37 ± 0.41	55.09 ± 0.63
Moraine MK-B4									
MK-10	32.85	BVOI	1.903	23.87 ± 0.32	8.39 ± 0.35	10.04 ± 0.18	0.39%	9.55 ± 0.40	54.90 ± 0.61
Moraine MK-B5									
MK-07	35.66	BVOG	1.902	25.70 ± 0.35	8.97 ± 0.21	8.37 ± 0.15	0.37%	9.30 ± 0.22	53.36 ± 0.66
Moraine MK-C1									
MK-04	35.26	BVOR	1.844	35.85 ± 0.57	6.95 ± 0.29	5.82 ± 0.12	0.67%	6.78 ± 0.28	55.24 ± 0.71
Moraine MK-C2									
MK-01	35.09	BVOP	1.852	24.86 ± 0.42	7.24 ± 0.31	9.27 ± 0.20	0.61%	7.45 ± 0.32	55.90 ± 0.63
MK-01a (<i>outlier</i>)	36.72	BVOK	1.859	22.36 ± 0.32	12.71 ± 0.44	9.55 ± 0.18	0.26%	12.79 ± 0.45	54.80 ± 0.54
Moraine MK-C3									
MK-02	35.49	BVOE	1.814	36.30 ± 0.51	2.144 ± 0.063	4.954 ± 0.099	1.69%	2.021 ± 0.061	53.38 ± 0.78
MK-02a	33.54	BVOL	1.872	38.1 ± 1.1	2.074 ± 0.089	5.08 ± 0.21	1.71%	2.125 ± 0.093	57.66 ± 0.69
MK-06	34.04	BVOF	1.897	30.26 ± 1.1	1.736 ± 0.091	7.018 ± 0.35	1.96%	1.81 ± 0.10	54.87 ± 0.67
Moraine MK-C4									
MK-09	31.06	BVOH	1.898	26.18 ± 0.37	0.431 ± 0.023	9.36 ± 0.17	7.76%	0.473 ± 0.030	54.95 ± 0.57

					Total number of Cl atoms (10^{15})	Total number of ^{36}Cl atoms (10^3)
blanc Hera	BVOO	1.882	267 ± 11	0.049 ± 0.011	359 ± 21	160 ± 36
blanc Athena	BVOD	1.812	148.6 ± 2.0	0.039 ± 0.011	741 ± 12	124 ± 35

Table 3. Individual ^{36}Cl sample and landform ages of Throne of Zeus (TZ) and Megala Kazania (MK) cirques. Ages are shown without denudation and snow correction as well as corrected with the snow correction factors of Table 1 and/or for a denudation rate of 5mm/ka. Please note that the ages only corrected for snow cover are those discussed in the text (see text for details). All uncertainties are reported at 1σ level. Uncertainties in the individual ages include the analytical and production rate errors, while italic numbers in parentheses are the analytical errors only. Uncertainties in the mean moraine ages include the standard deviation, analytical and production rate errors, while italic numbers in parentheses are the standard deviations only. Outlier samples are shown in italics.

No correction		Snow correction only		Erosion correction (5mm/ka)		Snow correction & erosion correction (5mm/ka)	
Sample & Landform ID	Age (ka)	Sample & Landform ID	Age (ka)	Sample & Landform ID	Age (ka)	Sample & Landform ID	Age (ka)
TZ-A	14.13 ± 2.03 (0.99)	TZ-A	14.86 ± 2.15 (1.08)	TZ-A	14.58 ± 2.19 (1.20)	TZ-A	15.42 ± 2.33 (1.31)
TZ-03	14.83 ± 1.88 (1.04)	TZ-03	15.62 ± 1.97 (1.10)	TZ-03	15.44 ± 1.95 (1.09)	TZ-03	16.35 ± 2.06 (1.15)
TZ-04	13.43 ± 1.67 (0.93)	TZ-04	14.09 ± 1.74 (0.97)	TZ-04	13.73 ± 1.71 (0.95)	TZ-04	14.50 ± 1.80 (1.00)
TZ-B	12.54 ± 1.76 (0.75)	TZ-B	13.23 ± 1.86 (0.82)	TZ-B	13.02 ± 1.88 (0.91)	TZ-B	13.80 ± 2.00 (1.00)
TZ-01	13.07 ± 1.67 (0.91)	TZ-01	13.82 ± 1.76 (0.97)	TZ-01	13.66 ± 1.76 (0.96)	TZ-01	14.50 ± 1.85 (1.02)
TZ-02	12.01 ± 1.51 (0.83)	TZ-02	12.65 ± 1.58 (0.88)	TZ-02	12.37 ± 1.56 (0.86)	TZ-02	13.10 ± 1.64 (0.91)
MK-B1	13.21 ± 2.67 (2.09)	MK-B1	13.94 ± 2.80 (2.19)	MK-B1	13.75 ± 2.80 (2.20)	MK-B1	14.58 ± 2.95 (2.31)
MK-11	11.73 ± 1.47 (0.85)	MK-11	12.39 ± 1.54 (0.90)	MK-11	12.20 ± 1.52 (0.88)	MK-11	12.94 ± 1.61 (0.94)
MK-12	14.69 ± 1.86 (1.02)	MK-12	15.49 ± 1.95 (1.08)	MK-12	15.31 ± 1.93 (1.06)	MK-12	16.22 ± 2.04 (1.13)
<i>MK-13</i>	<i>8.13 ± 1.04 (0.59)</i>	<i>MK-13</i>	<i>8.58 ± 1.10 (0.62)</i>	<i>MK-13</i>	<i>8.33 ± 1.07 (0.60)</i>	<i>MK-13</i>	<i>8.83 ± 1.13 (0.64)</i>
MK-B2	12.77 ± 2.17 (1.44)	MK-B2	13.49 ± 2.28 (1.51)	MK-B2	13.32 ± 2.27 (1.52)	MK-B2	14.13 ± 2.40 (1.60)
MK-14	13.79 ± 1.75 (0.96)	MK-14	14.56 ± 1.84 (1.01)	MK-14	14.40 ± 1.83 (1.00)	MK-14	15.27 ± 1.93 (1.06)
MK-15	11.75 ± 1.50 (0.81)	MK-15	12.42 ± 1.58 (0.86)	MK-15	12.25 ± 1.56 (0.85)	MK-15	13.00 ± 1.65 (0.90)
MK-B3	12.12 ± 1.90 (1.09)	MK-B3	13.32 ± 2.09 (1.20)	MK-B3	12.60 ± 2.00 (1.17)	MK-B3	13.92 ± 2.21 (1.31)
MK-03	11.34 ± 1.46 (0.82)	MK-03	12.47 ± 1.60 (0.90)	MK-03	11.77 ± 1.51 (0.85)	MK-03	13.00 ± 1.66 (0.94)
MK-05	12.89 ± 1.66 (0.93)	MK-05	14.17 ± 1.81 (1.02)	MK-05	13.43 ± 1.72 (0.97)	MK-05	14.85 ± 1.90 (1.07)
MK-B4	11.33 ± 1.42 (0.83)	MK-B4	11.97 ± 1.50 (0.90)	MK-B4	11.73 ± 1.47 (0.86)	MK-B4	12.43 ± 1.56 (0.92)
MK-10	11.33 ± 1.42 (0.83)	MK-10	11.97 ± 1.50 (0.90)	MK-10	11.73 ± 1.47 (0.86)	MK-10	12.43 ± 1.56 (0.92)
MK-B5	11.42 ± 1.42 (0.74)	MK-B5	12.06 ± 1.49 (0.78)	MK-B5	11.84 ± 1.47 (0.76)	MK-B5	12.55 ± 1.55 (0.81)
MK-07	11.42 ± 1.42 (0.74)	MK-07	12.06 ± 1.49 (0.78)	MK-07	11.84 ± 1.47 (0.76)	MK-07	12.55 ± 1.55 (0.81)
MK-C1	8.77 ± 1.13 (0.64)	MK-C1	9.64 ± 1.24 (0.71)	MK-C1	9.03 ± 1.16 (0.66)	MK-C1	9.97 ± 1.28 (0.73)
MK-04	8.77 ± 1.13 (0.64)	MK-04	9.64 ± 1.24 (0.71)	MK-04	9.03 ± 1.16 (0.66)	MK-04	9.97 ± 1.28 (0.73)

MK-C2	8.76 ± 1.13 (0.64)	MK-C2	9.61 ± 1.23 (0.71)	MK-C2	9.00 ± 1.16 (0.66)	MK-C2	9.91 ± 1.27 (0.73)
MK-01	8.76 ± 1.13 (0.64)	MK-01	9.61 ± 1.23 (0.74)	MK-01	9.00 ± 1.16 (0.66)	MK-01	9.91 ± 1.27 (0.73)
<i>MK-01a</i>	<i>15.30 ± 1.95 (1.06)</i>	<i>MK-01a</i>	<i>16.81 ± 2.13 (1.17)</i>	<i>MK-01a</i>	<i>16.05 ± 2.03 (1.12)</i>	<i>MK-01a</i>	<i>17.75 ± 2.25 (1.24)</i>
MK-C3	2.37 ± 0.32 (0.12)	MK-C3	2.51 ± 0.34 (0.13)	MK-C3	2.39 ± 0.33 (0.12)	MK-C3	2.53 ± 0.35 (0.13)
MK-02	2.44 ± 0.30 (0.16)	MK-02	2.58 ± 0.32 (0.17)	MK-02	2.46 ± 0.31 (0.16)	MK-02	2.61 ± 0.32 (0.17)
MK-02a	2.44 ± 0.31 (0.18)	MK-02a	2.58 ± 0.33 (0.19)	MK-02a	2.46 ± 0.32 (0.18)	MK-02a	2.60 ± 0.33 (0.19)
MK-06	2.23 ± 0.29 (0.18)	MK-06	2.35 ± 0.31 (0.19)	MK-06	2.24 ± 0.30 (0.18)	MK-06	2.37 ± 0.31 (0.19)
MK-C4	0.58 ± 0.08 (0.05)	MK-C4	0.64 ± 0.09 (0.05)	MK-C4	0.58 ± 0.08 (0.05)	MK-C4	0.64 ± 0.09 (0.05)
MK-09	0.58 ± 0.08 (0.05)	MK-09	0.64 ± 0.09 (0.05)	MK-09	0.58 ± 0.08 (0.05)	MK-09	0.64 ± 0.09 (0.05)

4. Mount Olympus Late-glacial and Holocene glacial chronology

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4.1 Throne of Zeus (TZ) cirque

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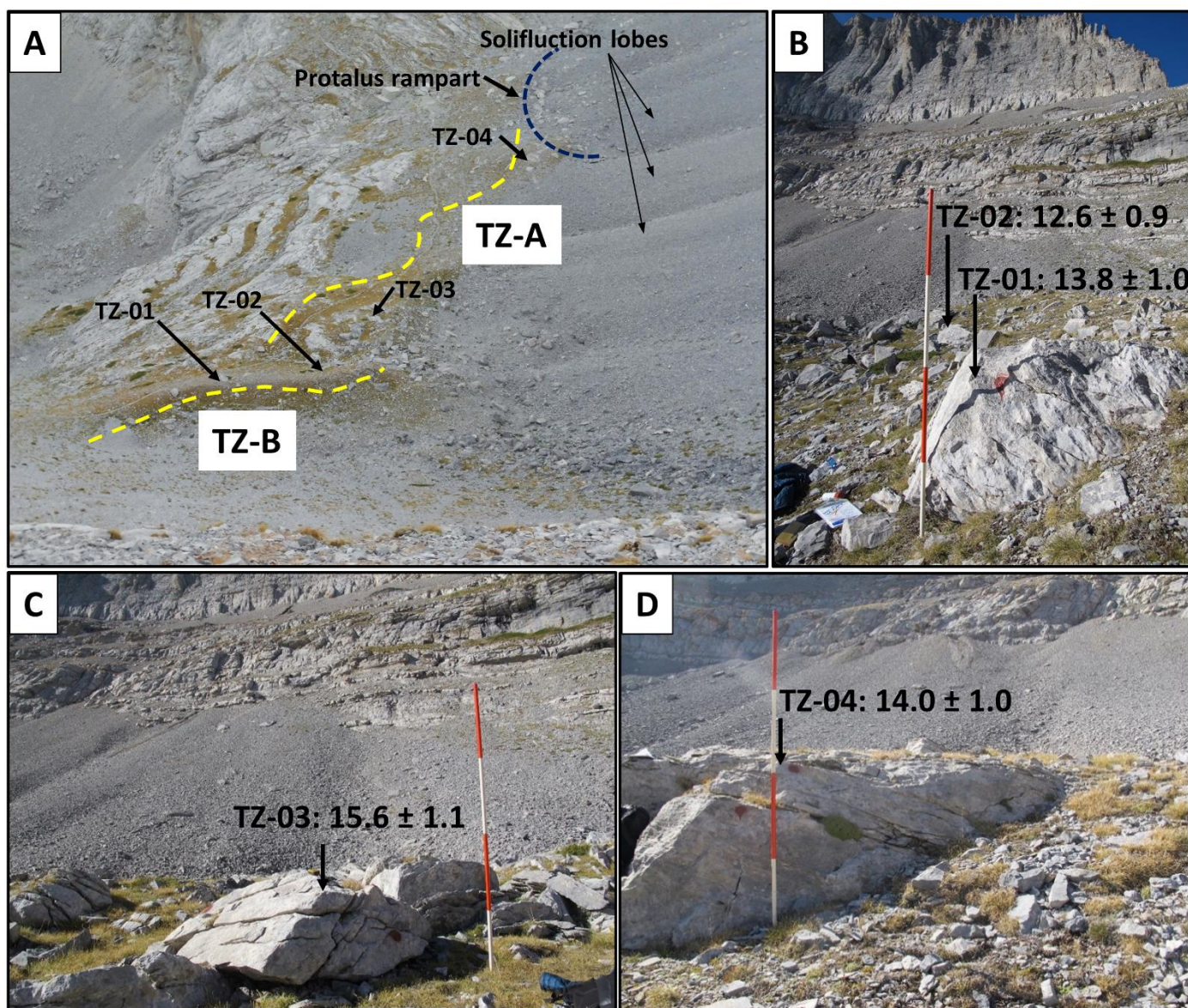
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The ^{36}Cl ages of the three boulders (TZ-01–03) and one bedrock sample (TZ-04) from TZ cirque, range between 15.6 ± 1.1 and 12.65 ± 0.88 ka (Table 3). Boulder sample TZ-03 (15.62 ± 1.1 ka) lies on the inner slope of frontal moraine TZ-A, 0.5m below its crest, while bedrock sample TZ-04 (14.08 ± 0.98 ka) was taken from a bench that outcrops along the inner side of TZ-A (Fig. 6C, D).



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Fig. 6. A) General view of Throne of Zeus cirque frontal moraine TZ-A and lateral moraine TZ-B, together with the sampled boulders and bedrock. B, C, and D) Sampled boulders TZ-01 and TZ-02, TZ-03 and TZ-04, with their respective calculated ages including analytical errors only.

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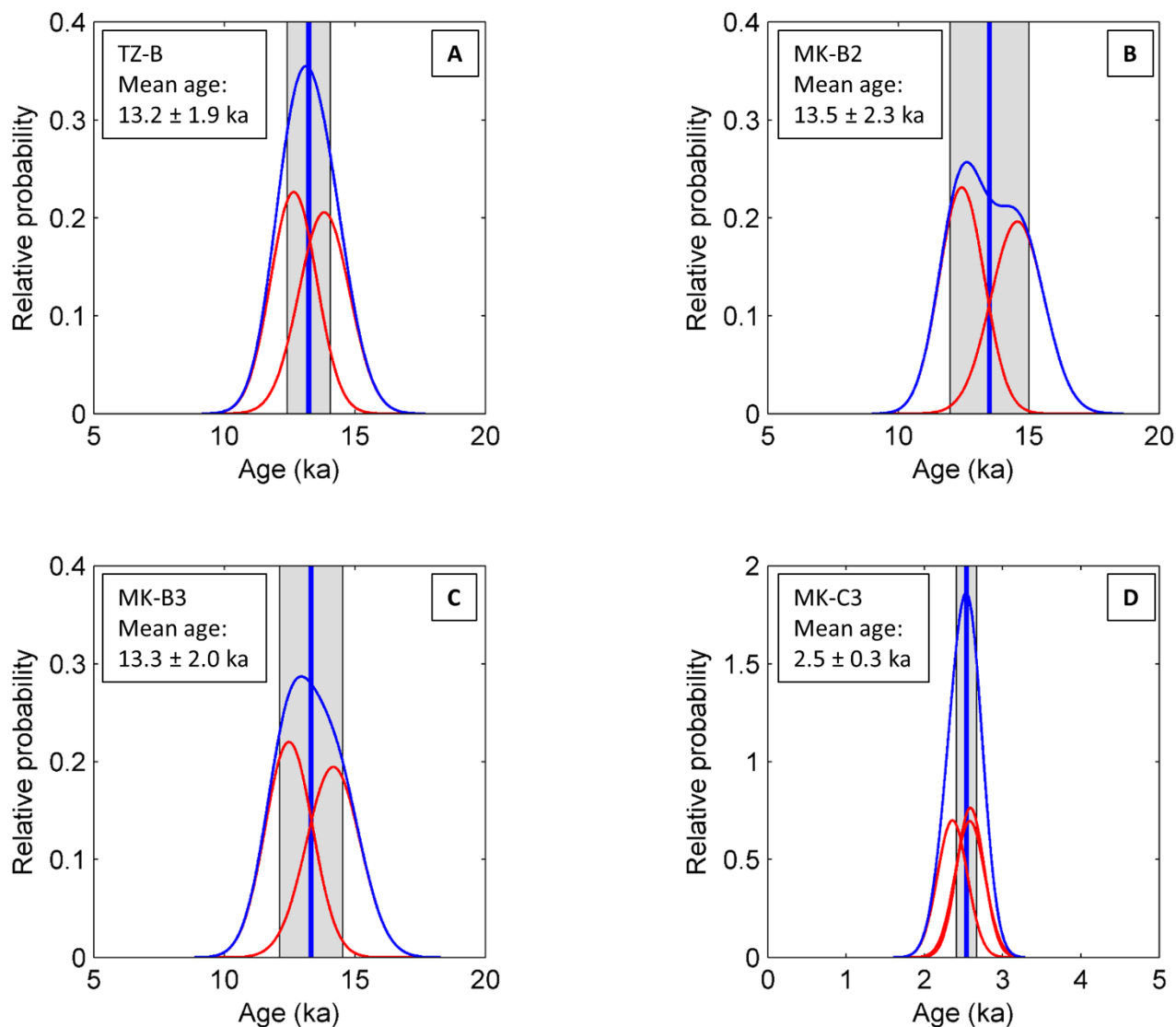
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The individual ages of samples TZ-03 and TZ-04 were not tested against the χ^2 criterion, because the age of boulder sample TZ-03 is considered to correspond to the moment of stabilization of the moraine, while bedrock sample TZ-04 was most likely exposed during the subsequent glacier retreat, as suggested by the chronological order of their ^{36}Cl ages. Samples TZ-01 (13.82 ± 0.97 ka) and TZ-02 (12.65 ± 0.88 ka) were obtained from flat-top parts of embedded boulders located on the outer slope of moraine TZ-B, 2m below the well-defined crest (Fig. 6A, B). They give a mean age of 13.2 ± 1.9 ka (Fig. 7A). The stratigraphical

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position of the geomorphologically distinct moraines TZ-A and TZ-B, imply a two-phase glacier behaviour of the TZ paleoglacier; a stabilization phase at 15.6 ± 1.1 ka that terminated around 14.0 ± 1.7 ka with the exposure of bedrock (sample TZ-04), and a subsequent re-advance or stagnation phase that resulted in the built-up and stabilization of moraine TZ-B at 13.2 ± 1.9 ka (Fig. 6A).



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Fig. 7. Probability density function plots of the TZ and MK landforms (moraines and pronival rampart) with more than one ^{36}Cl age that comply with the χ^2 criterion test. Individual boulder ages (red Gaussian distributions) are shown with their analytical errors only. The landform age distribution is represented by the blue summary curve. Blue vertical lines and grey bands are the arithmetic means and standard deviations of the landform ages, while the arithmetic mean ages given in the upper left corner of each panel are shown with the full uncertainty (including standard deviation, analytical and production rate errors). Ages are corrected for snow cover.

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Megala Kazania (MK) cirque

The sixteen ($n=16$) ^{36}Cl boulder ages sampled in MK cirque range from 15.5 ± 1.1 ka to 0.64 ± 0.05 ka (Table 3), spanning the Late-glacial and Holocene. Moraine MK-B1 is a composite feature as evident from several soil horizons observed on its outer slope and on the outwash plain on its inner slope (Fig. 4, 5 and 8). Boulders resting on the crest (yellow dashed line in Fig. 8) showed high degree of fragmentation and signs of block rotation and were therefore not suitable for sampling. Instead, three boulders (MK-11, MK-12, and MK-13) were sampled from different levels of the ~65 m high steep outer slope, i.e. 5m, 10m, and 2m below the crest, respectively (Fig. 4). Boulders MK-11 and MK-12 were in stable positions, embedded in the moraine's slope, thus minimizing the potential of block toppling from above, whereas boulder MK-13 was resting on the slope, in a rather instable position.



Fig. 8. A: Panoramic view of the Megala Kazania cirque, bounded by the 500m north wall of Skolio (2912m), together with the ages (analytical errors only) of moraines MK-B1, -B3, -B4, B5, C1, -C2, -C3, -C4 and protalus rampart MK-B2. The photo was taken on June 30th 2012. As in Fig. 5, yellow and light blue dashed lines correspond to landforms of morainic complexes MK-B and MK-C, respectively.

All three calculated ages of 12.40 ± 0.90 ka (MK-11), 15.5 ± 1.1 ka (MK-12) and $8.6x \pm 0.6x$ ka (MK-13) fail to meet the χ^2 criterion test. Sample MK-13 provided the youngest age ($8.6x \pm 0.6x$ ka) and was rejected as an outlier, because the mean ages of the moraines that are located further inboard were older, as presented below (Fig. 9C). This boulder was most likely affected by post-depositional movement, as can also be suspected from its instable position. Concerning the other two samples from moraine MK-B1, boulder MK-12 (15.5 ± 1.1 ka) is embedded in a stratigraphically lower position than MK-11 (12.40 ± 0.90 ka), consistent with their progressively younger ^{36}Cl ages (Fig. 4, Tables 2 and 3). Therefore, we attribute their age offset primarily to different phases of moraine formation resulting from recurrent glacier advances of equal extent. We consider the oldest age (sample MK-12: 15.5 ± 1.1 ka) as the most relevant for the timing of moraine MK-B1 deposition, which along with the age of sample TZ-03 (15.6 ± 1.1 ka), mark the beginning of an early Late-glacial moraine stabilization phase on Mount Olympus.

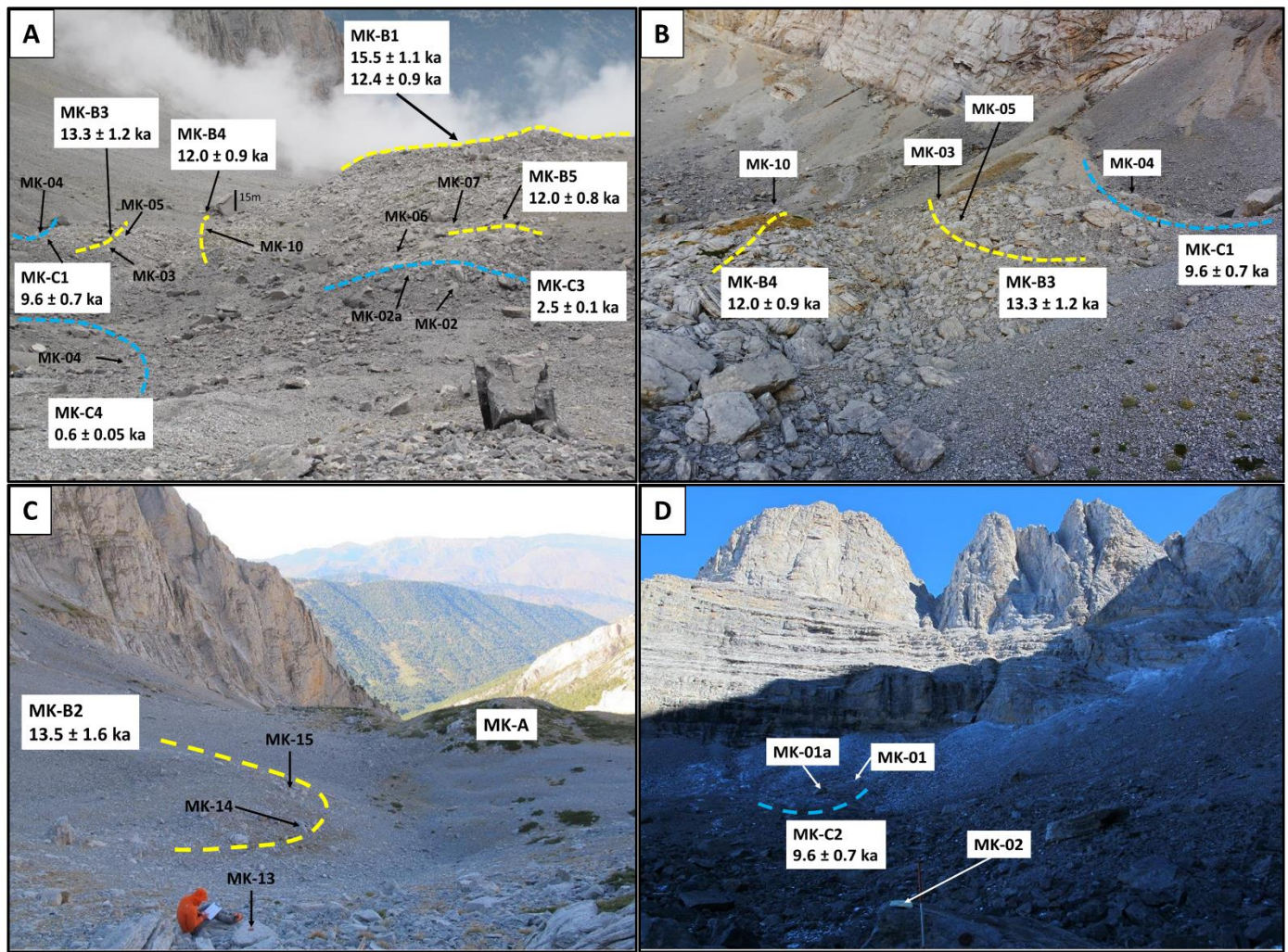


Fig. 9. Field photos of sampled boulders and landforms in MK cirque. A) General view in the down-valley direction of frontal moraines MK-B1 and MK-B3, hummocky moraines MK-B4 and MK-B5, together with the late Holocene moraines MK-C3 and MK-C4, with the locations of individual samples. B) Closer view in the upvalley direction of moraine MK-B3, together with nested moraine MK-C1 and hummocky moraine MK-B4. C) Moraines MK-A and MK-B2 located stratigraphically below frontal moraine MK-B1 (outlier sample MK-13 is also indicated). The photo was taken from the crest of moraine MK-B1 with view in the down-valley direction D) Early Holocene moraine MK-C2, situated close to the cirque headwall at the apex of the outwash plain. The photo was taken from the crest of the late Holocene moraine MK-C3. View in upvalley direction.

Along its western end (west of the cirque centre line, Fig. 5), moraine MK-B1 is dissected by push moraine MK-B2 (Figs 5, 8 and 9C). Two boulder samples, embedded in the outer slope of MK-B2, yield ages of 14.6 ± 1.0 ka (MK-14) and 12.43 ± 0.86 ka (MK-15), with a mean age of 13.5 ± 1.5 ka (Fig. 7B).

Further inboard of MK-B1 and closer to the cirque cliffs (southwest of the cirque centerline and below the high walls of Skolio), two boulders were sampled along the outer crest of moraine MK-B3 (Figs. 5 and 9B), yielding ages of 12.47 ± 0.91 ka (MK-03) and 14.2 ± 1.0 ka (MK-05) with a mean age of 13.3 ± 1.2 ka (Fig. 7). The formation of moraines MK-B2 and MK-B3, is related to excess of wind-blown snow accumulation deposited during a period with enhanced aeolian activity from the cliffs above.

Also inboard of frontal moraine MK-B1 but closer to the cirque center, the apron of large-sized (> 0.5 m) blocks with high degree of fragmentation (Fig. 5, 8 and 9A) is interpreted as englacial or supraglacial debris, deposited as the MK paleoglacier was retreating. This may have occurred either soon after the early Late-glacial phase of Mount Olympus moraine stabilization (~ 15.5 ka ago), or later after a second phase of Late-glacial glacier activity (~ 12.5 ka), which was probably associated to very cold conditions causing the fragmentation of these boulders. Boulder MK-10 on MK-B4 yielded an age of 11.97 ± 0.9 ka, and boulder MK-07 on MK-B5 gave an identical age of 12.06 ± 0.78 ka. We consider the ages of these hummocky moraines as representing the end of the second Late-glacial phase of glacial activity on Mount Olympus. The subsequent stages of glacier dynamics in MK cirque correspond to the deposition of moraine group MK-C and are of Holocene age. Blocky moraine MK-C1, which is nested within moraine MK-B3 close to the southwestern part of the cliff (Figs 5, 8, 9A and 9B), was dated to 9.64 ± 0.71 ka based on boulder sample MK-04. Two boulder samples were collected from another blocky moraine, MK-C2, which is

480 located in an approximately symmetrical position compared to MK-C1 relative to the cirque center line (Fig.
481 5), and yielded ages of 9.62 ± 0.71 ka (MK-01) and 16.8 ± 1.2 ka (MK-01a) (Table 3). The old age of
482 sample MK-01a is incompatible with the rest of the moraine chronology in MK glacial cirque, most likely
483 due to inherited ^{36}Cl concentrations from previous exposure periods, and it was thus considered an outlier
484 and discarded from the discussion. The age of boulder MK-01 is identical with that of boulder MK-04 from
485 moraine MK-C1, implying that both moraines MK-C1 and -C2 were formed at the same time (~ 9.6 ka ago)
486 as a result of a common forcing mechanism that restricted the ice extension along the perimeter of the cirque
487 close to the bounding cliffs. Their deposition is therefore attributed to an early Holocene glacier re-advance
488 or standstill, as the MK paleoglacier was retreating from its larger late Late-glacial extent. From our
489 chronology we cannot draw further conclusions whether following this early Holocene glacier extent, MK
490 paleoglacier persisted in sheltered locations close to the cirque headwalls, or disappeared entirely during the
491 mid-Holocene. The next phase of glacial activity on Mount Olympus is evident from the central frontal
492 moraine MK-C3 (Fig. 5, 8, 9A), dated with three boulders samples to 2.59 ± 0.17 ka (MK-02), 2.58 ± 0.19
493 ka (MK-02a), 2.36 ± 0.19 ka (MK-06), with a mean age of 2.51 ± 0.13 ka (Fig. 7). The corresponding late
494 Holocene glacier likely occupied a similar or slightly larger area in relation to the early Holocene standstill
495 boundaries (Figs 10F, G). Finally, blocky moraine MK-C4 is confined close to the southern cirque margins,
496 100m below the terminus of the present-day permanent snowfields lower boundary (Styllas, 2016). The
497 single date of 0.64 ± 0.05 ka (boulder MK-09) from this moraine places its formation within the early part of
498 the Little Ice Age. It is also likely that the formation of MK-C4 in such protected location can be attributed
499 to a large degree to the local topoclimatic factors, the most important being the accumulation of wind-blown
500 and avalanching snow during the winter and shading from the cliffs above during the summer (Fig. 5, 8,
501 9A).

502 Noteworthy is the absence of landforms of similar age in the Throne of Zeus cirque, which highlights the
503 importance of orientation among other topoclimatic factors (height of headwalls, wind-blown snow, shading
504 etc.) for the formation and evolution of Mediterranean cirque glaciers.

505 5. The tempo of Mount Olympus glacial history in relation to regional glacial, lacustrine and marine 506 environments

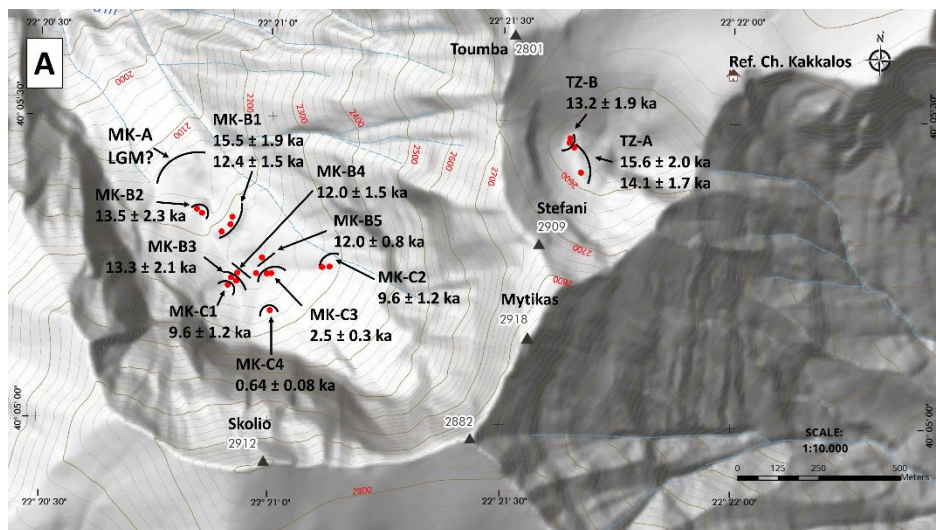
508 Based on our calculated landform ages and geomorphological investigations in the Throne of Zeus (TZ) and
509 Megala Kazania (MK) cirques, we propose a chronological sequence of Mount Olympus glacial history, and
510 compare it with the existing SED studies from the northeast Mediterranean mountains and with paleoclimate
511 proxies from regional terrestrial and marine environments. In this section, we show the individual ^{36}Cl ages
512 and arithmetic mean ages with their full uncertainties (including production rates, see section 3.2), to allow
513 for comparison with chronologies based on other dating methods (Table 3, Fig. 7). We note that the error
514 ranges of the landform ages permit us to only tentatively relate our findings to specific paleoclimate periods,
515 such as the Oldest Dryas ($17.5 - 14.7$ ka, Rasmussen et al., 2006) and the Younger Dryas ($12.7 - 11.5$ ka -
516 Alley, 2000), which are temporally well-constrained from regional continuous high resolution paleoclimatic
517 records, e.g. from the records of oxygen isotope changes in Greenland ice cores. We rather assign the
518 distinct phases of glacial activity on Mount Olympus to the early, middle and late stages of the Late-glacial
519 and to the early ($11.7 - 8.2$ ka) and late ($4.2 - 0$ ka) Holocene (Walker et al., 2012). However, it must also
520 be noted that the chronological order of our ^{36}Cl mean ages in both glacial cirques is in very good agreement
521 with the landform stratigraphy, thus strongly supporting the general temporal sequence of glacier
522 fluctuations that we infer from our ^{36}Cl chronology.

523 5.1 The Late-glacial and Holocene glacial chronology of the northeast Mediterranean Mountains

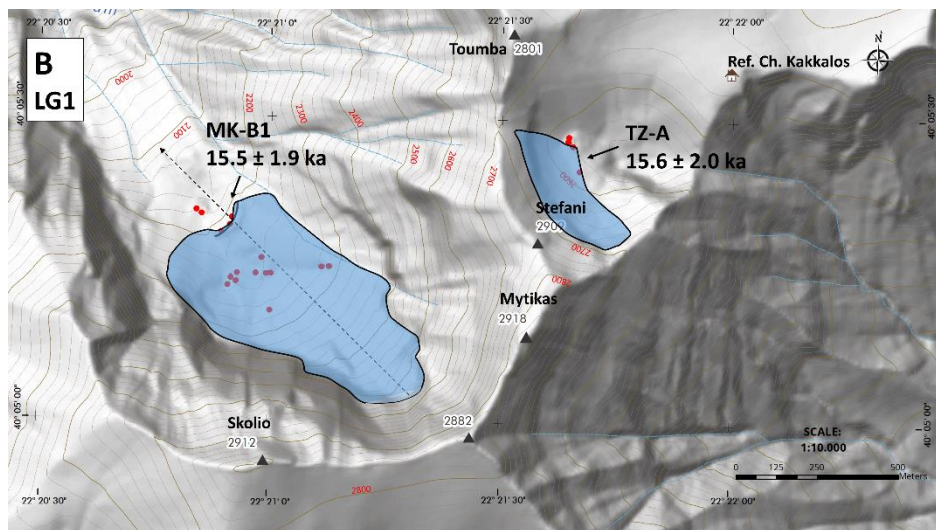
524 The first phase of glacial activity within the TZ and MK cirques with respective ELAs at $\sim 2600\text{m}$
525 and $\sim 2200\text{m}$, is derived from the individual ages of samples TZ-03 (15.6 ± 2.0 ka) and MK-12 (15.5 ± 2.0
526 ka), which are considered to represent the minimum ages of stabilization of frontal moraines TZ-A and MK-
527 B1, respectively (Fig. 10A, B). By taking into account the small surface area and the altitudinal extent of the
528 cirques under consideration, we assume no significant differences between the local ELA and the glacier
529 termini. In the following discussion, we refer to this time as the “early Late-glacial stage 1” (LG1). The LG1
530 timing of moraine stabilization is in good agreement with ^{10}Be ages of glacial landforms from Šara Range
531 and Mount Pelister (SW Balkan mountains) and from Mount Uludağ’s east Karagol valley and West Ski
532 Area in NW Turkey (see Fig. 1 for locations), where glacier advances with termini between 2000m and
533

534 2200m occurred no later than 16.4 ± 1.3 ka, 15.2 ± 0.85 ka, 15.9 ± 1.1 ka and 15.2 ± 1.0 ka respectively
 535 (Fig. 11) (Kuhlemann et al., 2009, Ribolini et al., 2017, Akçar et al., 2014, Zahno et al., 2010). Note that for
 536 comparison, the single-boulder age from Šara Range (16.4 ± 1.3 ka, Kuhlemann et al., 2009) was
 537 recalculated using the same calculator and parameters as in Ribolini et al. (2017) and Akçar et al. (2014), i.e.
 538 the online CRONUS-Earth calculator version 2.2 with the NE North American production rate and the time-
 539 dependent “Lm” scaling scheme (Balco et al. 2008, 2009). The concentration of this sample measured at the
 540 ETH tandem facility in Zürich relative to laboratory standard S555 (Kubik and Christl, 2010) was multiplied
 541 by 0.9124 to normalize to the 07KNSTD standard (Akçar et al., 2011). The ^{10}Be ages by Zahno et al. (2010)
 542 had already been recalculated accordingly, by Akçar et al. (2014).

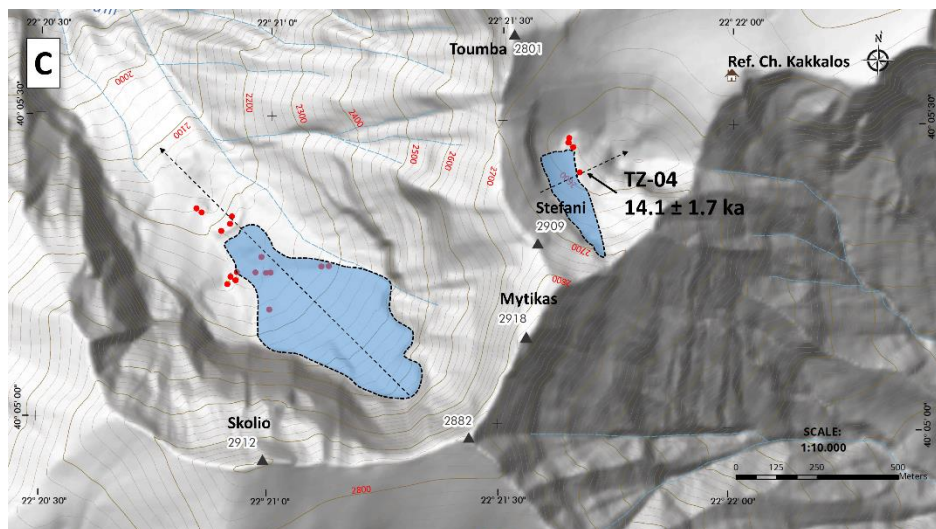
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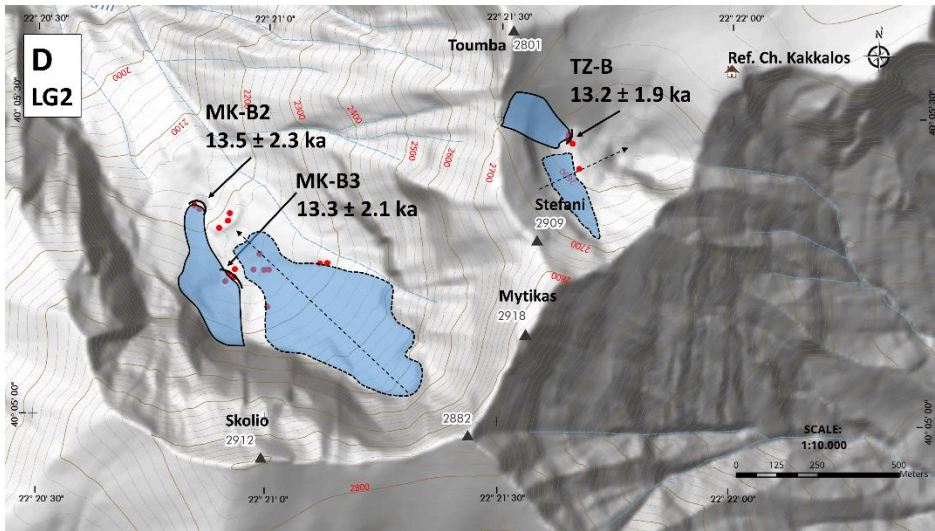
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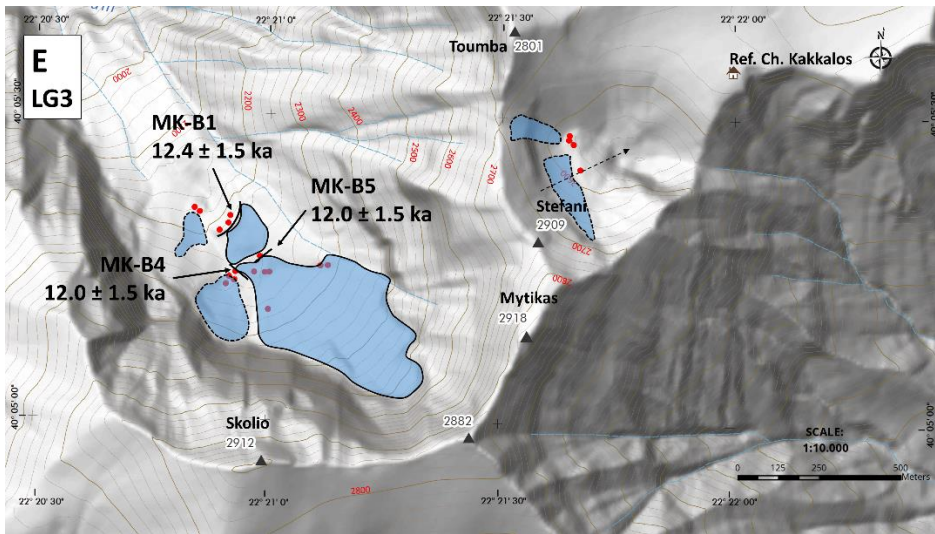
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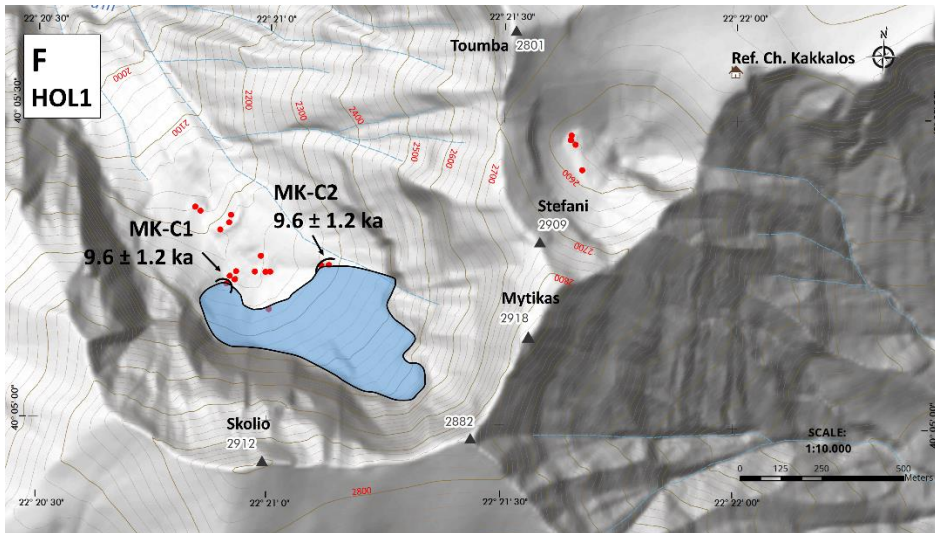
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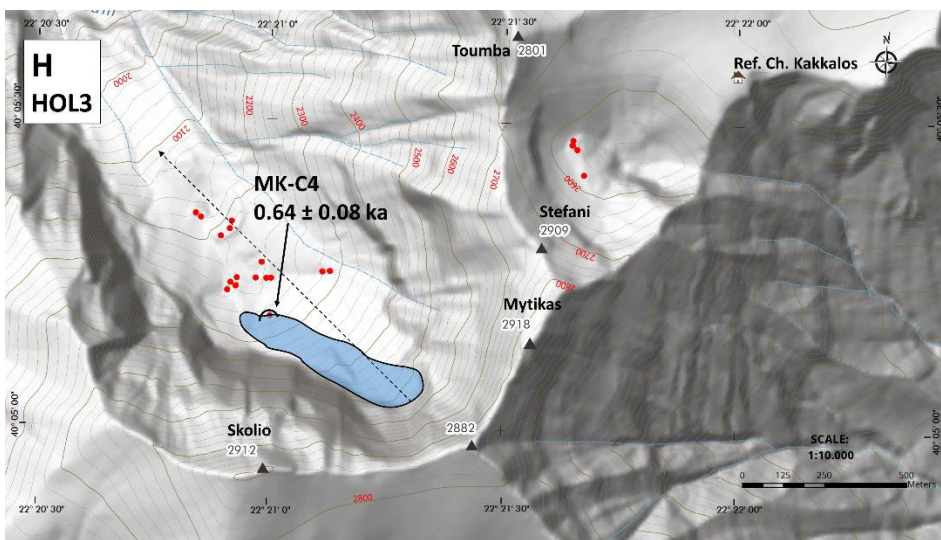
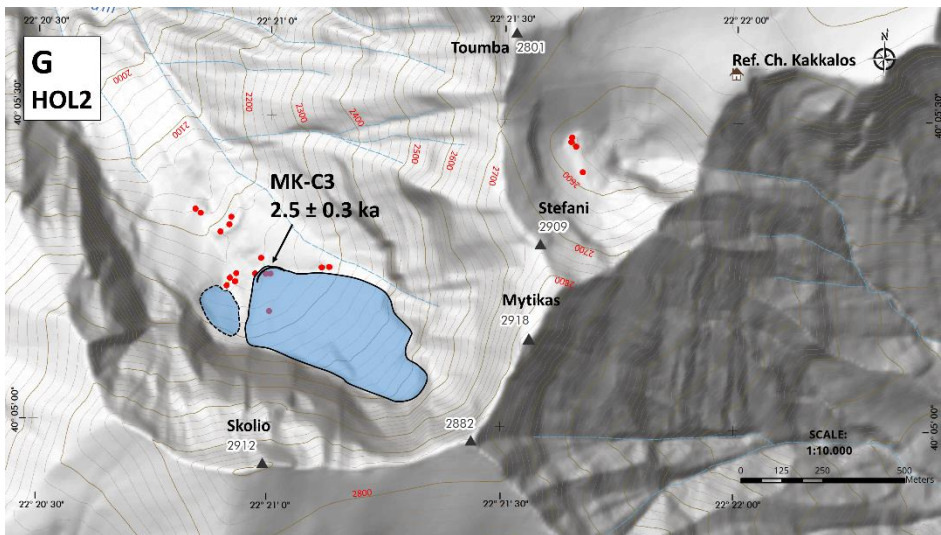


Fig. 10. Reconstructions of the Late-glacial and Holocene evolution stages of TZ and MK paleoglaciers, with the respective moraines and/or individual samples that delimit the glacier extensions of each stage. A) Dated landforms with mean ages (arithmetic mean) and individual samples ages, where only one sample was used to date the landform, or where the χ^2 criterion test failed. All ages include the full range (analytical and production rate) errors. Red dots correspond to the individual samples and black curved lines to the individual moraines shown in Fig. 2. In panels B to H, blue areas delimited by dashed lines correspond to approximate spatial boundaries of the paleoglaciers, tentatively defined by the analyses of this study. B) Early Late-glacial (15.5 ± 2.0 ka) stabilization phase (LG1) of moraines TZ-A and MK-B1. C) Early Late-glacial (14.1 ± 1.7 ka) retreat phase. D) Mid Late-glacial stabilization phase (LG2) of moraines TZ-B and MK-B3 and push moraine MK-B2, under windier conditions (13.5 ± 2.0 ka). E) Late Late-glacial (LG3) glacier shrinking under cold conditions with deposition of hummocky moraines MK-B4 and -B5, related to the Younger Dryas. F) Early Holocene glacier standstill phase (HOL1) with exposure of moraines MK-C1 and -C2 at 9.6 ± 1.2 ka. G) Late Holocene stabilization phase (HOL2) of moraine MK-C3 (2.5 ± 0.3 ka). H) Little Ice Age glacier expansion (HOL3) at 0.64 ± 0.08 ka.

On Mount Olympus, the temporal and spatial extents of TZ and MK paleoglaciers shortly after the LG1 moraine stabilization phase, cannot be defined with certainty from the existing data. The LG1 glacier extension phase was most likely followed by a gradual deglaciation, as manifested by the exposure age of 14.1 ± 1.7 ka of bedrock sample TZ-04. (Fig. 10C).

Recently, Gromig et al., (2017) used ^{36}Cl to date the stabilization of five limestone boulders on a moraine in the Galicica Mountains in SW Balkan Peninsular (Fig. 1), and related its formation to the Younger Dryas based on the mean moraine age of 12.0 ± 0.6 ka. The mean moraine age was corrected for an erosion rate of 5 mm/ka and for a snow cover factor of 0.98. The age calculations were done using a ^{36}Cl production rate for spallation of Ca (Marrero et al., 2016) higher by ~20% than the one we use in our study, and the uncertainty in mean age does not include the production rate errors. For the sake of comparison, we recalculated the five boulder ages using the same ^{36}Cl production and scaling parameters as applied for our samples, and taking into account the erosion and snow corrections locally suggested in Gromig et al. (2017) (5 mm/ka erosion; snow cover factor of 0.98). This resulted in a mean moraine age and full error of $14.0 \pm$

585 1.8 ka. We suggest that this latter age is in better agreement with other evidence of glacial behavior in this
586 region based on the following observations. The dated moraine in the Galicica Mountains is located at an
587 elevation of 2030m and is the largest and second-to-outmost one in a series of several moraines. On Mount
588 Pelister, located 30km to the east across Lake Prespa (Fig. 1), the most prominent of a series of moraines
589 located at an elevation of 2230m, composed of quartz-bearing lithologies, was dated with ^{10}Be (for which
590 the production rate is well constrained) to a mean age of 15.2 ± 0.85 . The formation of this moraine was
591 assigned to the Oldest Dryas glacial advance (Ribolini et al., 2017). Given, 1) the proximity between the two
592 sites in the Galicica Mountains and on Mount Pelister, 2) the similar north-eastern orientation of both glacial
593 cirques and 3) the ~200 m higher elevation of the Oldest Dryas moraine on Mount Pelister, we suggest that a
594 Younger Dryas age for the moraine in the Galicica Mountains is rather unlikely. We further propose that the
595 moraine sequences in both the Galicica Mountains (Gromig et al., 2017) and on Mount Pelister, like in the
596 case of Megala Kazania, can be related to a series of Late-glacial successive advances of similar extent.
597

598 The LG1 glacial phase is different from a preceding glacial phase recorded in Rila Mountains (~18–
599 16 ka), and which has been ascribed to a late stage of the LGM (Kuhlemann et al., 2013). On Mount
600 Olympus, this pre-LG1 glacial phase is likely represented by the stabilization of moraine MK-A (Fig. 9C
601 and 10A). We thus conclude that moraine stabilization during LG1 glacial phase in the northeast
602 Mediterranean mountains occurred at ~15.5 ka with an ELA at ~2200m in the north/northeast facing cirques
603 of Šara Range, Galicica Mountains, Mount Pelister, Mount Uludağ and MK, and at ~2600m in the east
604 facing TZ cirque and terminated as the glaciers began to retreat at ~14.0 ka (Fig. 11). Within uncertainties,
605 the LG1 glacial phase can be tentatively ascribed to a late stage of the Oldest Dryas (Fig. 11), which ended
606 ~14.7 ka ago (Rasmussen et al., 2006). This is in line with the postulation by Hughes et al (2003, 2006a) that
607 optimal conditions for Mediterranean glaciations occurred in intermediate periods between major stadials
608 and interstadials, as during major stadials reduced moisture availability was not sufficient to promote
609 glaciation.
610

611 The glacier retreat phase at ~14.0 ka (Fig. 10C) was followed by a return to glacial conditions with the
612 stabilization of moraines MK-B2 (13.5 ± 1.6 ka.), MK-B3 (13.3 ± 1.2 ka) and TZ-B (13.2 ± 1.9 ka),
613 suggesting a common forcing mechanism (Fig 10D). The stabilization of the three moraines occurred during
614 a mid-Late-Glacial period of climatic change, which we define here as the Late-glacial stage 2 (LG2), (Fig.
615 11). The stabilization of moraines TZ-B, below the saddle separating the summits of Stefani (2909m) and
616 Toumba (2801m), and of MK-B2 and -B3 below the 500m north facing headwall of Skolio (2912m) (Fig.
617 2), can be attributed to excess accumulation of wind-blown and avalanching snow related to wind activity
618 from a general western direction (Figs. 3, 10D). The increased deposition of wind-blown snow likely
619 occurred in marginal conditions for glaciation, as suggested by the regional warmer conditions around ~13–
620 14 ka that are recorded in numerous lacustrine and marine sedimentary records along the northeast
621 Mediterranean (see section 5.2). The warmer and windier conditions during LG2, terminated with another
622 return to glacier friendly conditions during the late Late-glacial, as suggested by sample MK-11 (12.4 ± 1.5
623 ka), which might denote a post-depositional reactivation of frontal moraine MK-B1. The third phase of
624 glacial activity during the Late-glacial stage 3 (LG3), is associated with the stabilization of hummocky
625 moraines MK-B4 and -B5 (Figs 10E, 11). Their positions, 150m inboard and 20m higher relative to frontal
626 moraine MK-B1 and closer to the center of the cirque, suggest that LG3 was characterized by glacier decay.
627 Observations from the Svalbard region and studies of Younger Dryas moraines in Scotland relate the
628 formation of hummocky moraines to changes in the glacier thermal regime from temperate to polythermal,
629 with warm-based ice in the interior and cold-based ice in the glaciers margins (Hambrey et al., 1997).
630 Therefore, it is likely that the formation and exposure of the hummocky moraines MK-B4 and MK-B5 at
631 12.0 ± 1.5 ka corresponds to a significant drop in air temperatures, also evident from the cryogenic features
632 within the sedimentary sequence of Theopetra cave (Fig. 1, Karkanis, 2001). This implies that conditions
633 were cold enough and moisture availability was considerably low during LG3, resulting in an overall cold
634 and dry phase of glacier decay, which is most likely responsible for the intense fragmentation of the boulder
635 apron inboard of moraine MK-B1. Within dating uncertainties, we ascribe the LG3 glacial phase,
636 characterized by glacial decay to the Younger Dryas (Alley et al., 1997).
637

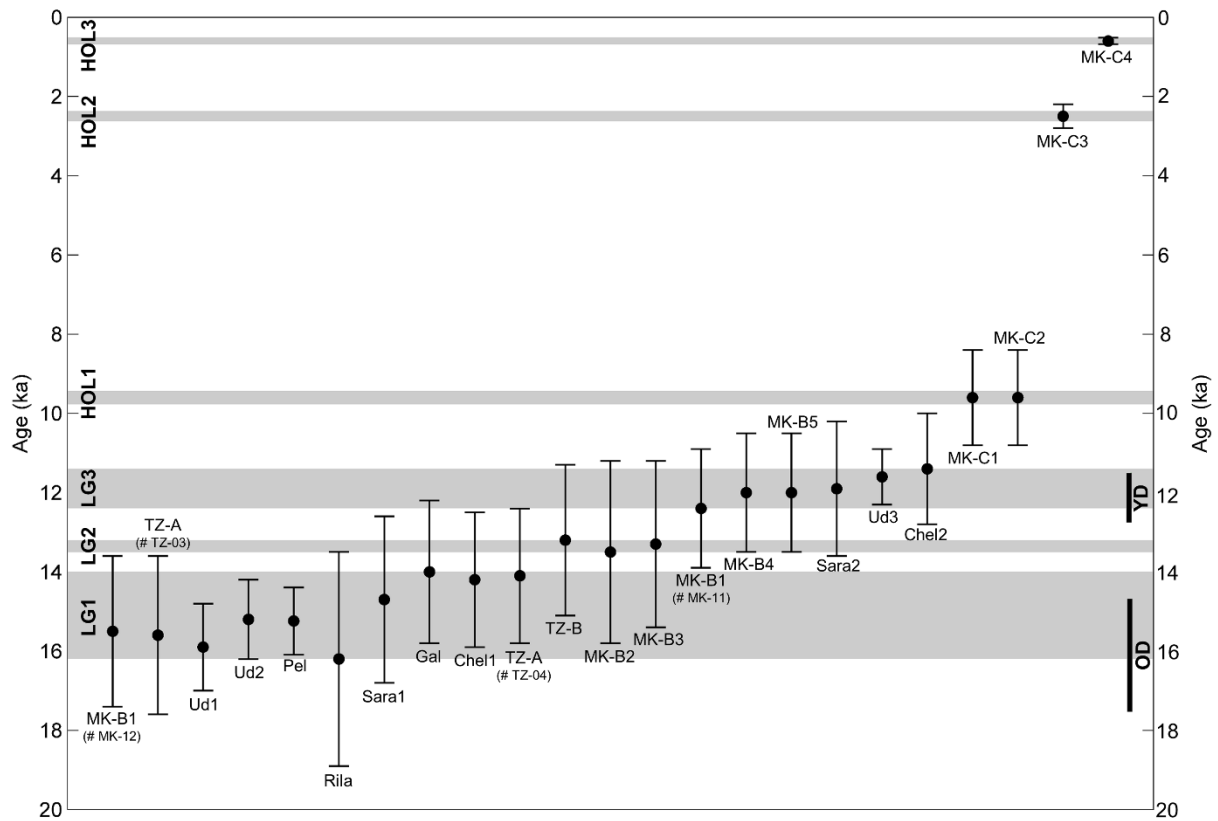


Fig. 11. The Late-glacial and Holocene glacial chronology of the northeast Mediterranean mountains based on SED of glacial landforms. Glacial phases (LG1-3, HOL1-3), resulted from the minimum and maximum mean ages of the dated landforms. Mount Olympus landforms MK-B1 and TZ-A are represented by two separate samples each (shown in parentheses), while the others are represented either by a single sample, or by the arithmetic mean of the independent ages that meet the χ^2 criterion (Fig. 7). Other landforms include SED from: Ud1: Uludağ Mountain Karagöl Valley (Akçar et al., 2014). Ud2: Uludağ Mountain West-Ski Area (Zahno et al., 2010). Pel: Mount Pelister (Ribolini et al., 2017). Rila: Rila Mountains (Kuhlemann et al., 2012). Sara1: Šara Range sample S16 (Kuhlemann et al., 2009). Gal: Galicica Mountains (Gromig et al., 2017), arithmetic mean of the recalculated ages with the same parameters of this study. Chel1, 2: Mount Chelmos samples from Kato Kambos valley (Pope et al., 2015), recalculated with the same parameters of this study. Sara2: Šara Range sample S7 (Kuhlemann et al., 2009). Ud3: Uludağ Mountain West-Ski Area (Zahno et al., 2010). The temporal boundaries of the Older Dryas (Rasmussen et al., 2006) and the Younger Dryas (Alley et al., 1997) periods, are shown.

Despite the ~350 m elevation difference between the frontal moraines MK-B1 in MK cirque (2225m) and TZ-A in TZ cirque (2580m), both moraines formed during the same period (LG1). This can be explained by the different topographical configurations of the two cirques. MK cirque has a NW orientation, is bounded by 500m high headwalls, receives higher amounts of windblown snow and has a longer period of shading. These geomorphological characteristics permit the formation of a larger and more protected glacier that extended to lower altitudes and occupied a surface area 3 times larger (1 km long, 0.5 km wide) compared to the TZ paleoglacier. Due to these topographic attributes, deglaciation of the MK paleoglacier was much slower than that of TZ paleoglacier. This becomes particular apparent from the fact that the glacier survived after the LG1-3 phases evident from the deposition of morainic complex MK-C, which corresponds to glacier extents during the early (11.7 – 8.2ka) and late (4.2 – 0ka) Holocene (Walker et al., 2012). Geomorphological evidence of Holocene glaciation are also present as glacier activity in Šara Range (Kuhlemann et al., 2009) and the Rila Mountains (Kuhlemann et al., 2013), but are either obscured and/or absent from Galicica Mountains, Mount Pelister and TZ cirque. We interpret the early Holocene blocky moraines MK-C1 and -C2 at 9.6 ± 1.2 ka, located close to the cirque headwalls, as representative of a glacier standstill or short readvances (HOL1, Fig. 11), during a general early Holocene phase of warming and glacier retreat (Fig. 10F). The late Holocene glacier expansions (Fig. 10G and H) with stabilization of moraines MK-C3 and MK-C4 at 2.5 ± 0.3 ka and 0.6 ± 0.08 ka (Fig. 11), respectively, may be explained by a return to wet winters and cool summers that promoted the last glaciation phases on Mount Olympus.

5.2 Correlations of Mount Olympus glacial chronology with regional terrestrial and marine records

In this section, we correlate our new glacial chronologies with selected paleoclimatic proxies, in order to investigate possible external and internal climatic forcing mechanisms responsible for the climate changes that are recorded by the glacial landforms on Mount Olympus. We emphasize again that these correlations must be considered tentative given the uncertainties in the cosmogenic nuclide ages.

Phase LG1 characterized by moraine stabilization within the TZ and MK cirques and subsequent glacier recession at ~14.0 ka (Fig. 11), coincides with a cycle of peak solar insolation, as expressed by the Optical Depth of Luminescence (ODL) in Duhlata Cave (Fig. 12A). Despite the fact that the theoretical solar insolation curves by Berger and Loutre (1991) are the most accredited in paleoclimate reconstructions, they explain about ½ of the paleoclimatic signal (e.g. Imbrie et al., 1993). For this reason we use here the ODL record from the proximal (280km to the northeast, Fig. 1) to Mount Olympus Duhlata Cave. The ODL is derived from the calcite speleothem luminescence of organic material, which also takes into account the solar luminosity and thus considered an indirect regional proxy of solar radiation (Stoykova et al., 2008). Superimposed on this peak solar insolation cycle, is the change of the regional climate from cold and dry to warmer and wetter, suggesting an *out-of-phase* behaviour. The cold and dry climate during LG1 is registered in the lacustrine records of Lakes Prespa, Ohrid and Tenaghi Philippon (Fig. 12C, D, E). The two former lakes are located 140km west of Mount Olympus at elevations of 690 and 850m, with their watersheds bounded by Galicica Mountains and Mount Pelister, whereas the latter is located 190km to the northeast of Mount Olympus at an elevation of 50m. The existence of steppic taxa such as *Artemisia* and *Chenopodiaceae*, in Lake Prespa (Aufgebauer et al., 2012) and in Tenaghi Phillipon (Fig. 12E) (Pross et al., 2015, Wulf et al., 2018), indicate regional dry conditions. This is further supported by the low CaCO₃ values in Lake Ohrid, which suggest low allochthonous carbonate sediment input from the catchment (Fig. 12D), likely due to reduced surface runoff (Vogel et al., 2010). Quantitative pollen-based temperature and precipitation reconstructions from Lake Maliq (located 20km south of Lake Ohrid and 10km west of Lake Prespa, Fig. 1), suggest that between 16–15 ka, the mean annual temperature ranged from -3 to 1°C and the mean annual precipitation was 400mm, whereas present values are 11.2°C and 790mm, respectively (Bordon et al., 2009). At lower elevations along the Aegean Sea, this period was characterized by relatively cool (14.5°C) Sea Surface Temperatures (SST) (Fig. 12F), by an expanse of *Pinus* (Fig. 12G) (Kothoff et al., 2011), by increased fluvial sediment inputs (Fig. 12I) and by coarser grain-sized sediments (increased silt fractions) on the bottom of the north Aegean Sea (Fig. 12J). This can be explained by the fact that vegetation was sparse along the upper parts of watersheds and glacial conditions produced coarser sediments containing higher amounts of silt. Despite the overall cold conditions in the higher (>800m) elevations, precipitation starvation combined with high solar insolation (Fig. 12A), are considered as the likely causes that restricted the expansion of glaciers within the cirques at elevations between 2000 and 2200m. The transition to warmer and wetter climatic conditions towards the later part of LG1, resulted in the gradual deglaciation of the northeast Mediterranean cirques, which together with increasing annual precipitation (Bordon et al. 2009) and SST's (Fig. 12F), contributed to an expansion of *Quercus* forests and to increasing fluvial inputs, with higher amounts of fine fractions in the Aegean Sea (Fig. 12J).

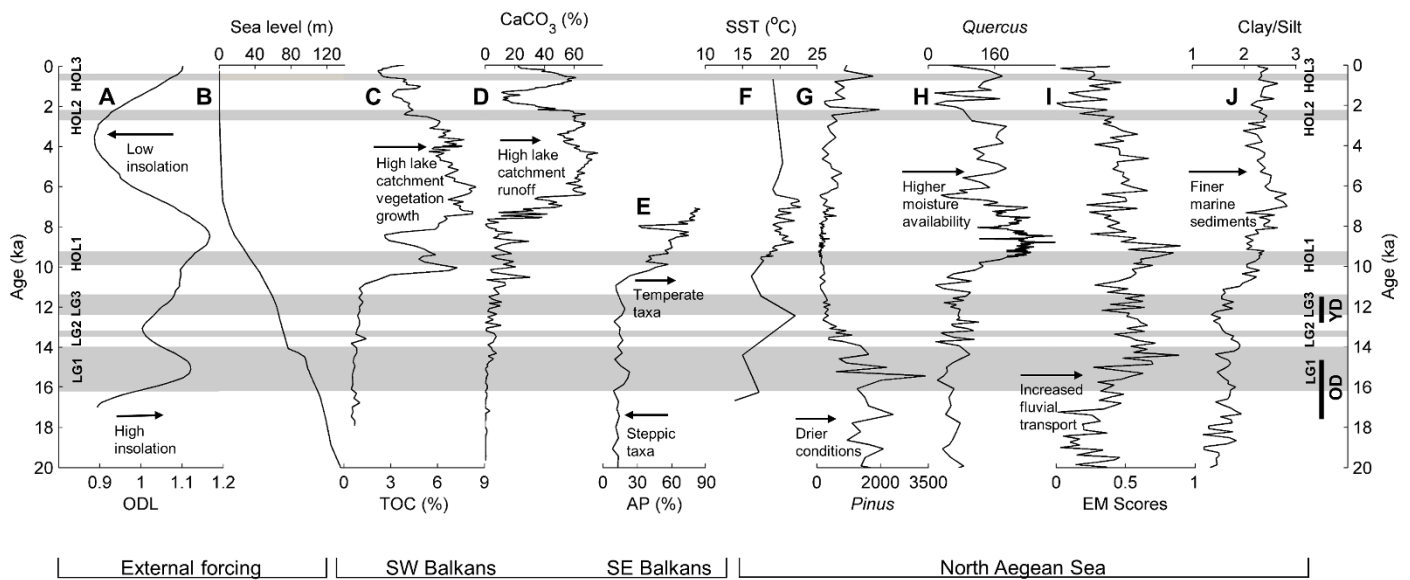


Fig. 12. Correlation of the Late-glacial and Holocene glacial phases (LG1-3 and HOL1-3) derived from the compilation of SED-based glacier chronologies from the northeast Mediterranean mountains (Fig. 11), with selected external forcing, terrestrial and marine proxies. A) Optical Depth of Luminescence, Duhlata Cave (Stoykova et al., 2008). B) Ice-volume equivalent sea-level function, expressed as m below present sea level (Lambeck et al., 2014). C) Total Organic Carbon from Prespa Lake (Aufgebauer et al., 2012). D) CaCO_3 concentrations from Lake Ohrid (Vogel et al., 2010). E) Tenaghi Philippon Arboreal Pollen (AP) concentrations (Wulf et al., 2018). F) North Aegean Sea SST's ($^{\circ}\text{C}$) (Core MNB-3, Gogou et al., 2007). G) *Pinus* and H) *Quercus* pollen counts from marine core SL-152 (Kothoff et al., 2008). I) End-member 2 (EM2) scores representative of fluvial input from marine core SL-148 (Hamman et al., 2008). J) Clay/Silt fraction of marine core SL-148 as a proxy of grain-size and glacial versus interglacial conditions (Ehrmann et al., 2007).

On Mount Olympus, the deglaciation that concluded LG1 was followed by glacial phase LG2 (Fig. 11), despite the rapid changes of the regional climate. Bordon et al. (2009) suggest the occurrence of an abrupt warming on the order of 10°C after LG1 (~15 ka), deduced from a sediment core record in Lake Maliq (elevation 690m) with mean annual temperatures $7\text{-}10^{\circ}\text{C}$. In north Aegean Sea a sharp increase in lipid biomarker-derived SST from 14.5°C to 22°C is recorded in core MNB 3 (Gogou et al, 2007, Fig. 1) between 14.5 and 12.5 ka (Fig. 12F). However, the stabilization of moraines MK-B2, -B3 and TZ-B during LG2, implies that despite the regional warming, the conditions on Mount Olympus remained favourable to glaciation, and we attribute this mainly to excess wind-blown snow accumulation and to solar insolation minima (Fig. 12A). The enhanced aeolian activity on Mount Olympus is also represented in Lakes Ohrid and Prespa sediment records by the increased amounts of sand in the bottom sediments, which have been related to subaqueous current activity under an intense wind stress field (Vogel et al., 2010). During the same time interval, the sea-level had risen from its LGM lowstand (-120 m) to a depth of -70 m (Fig 12F), and was distanced 35km from TZ and MK cirques. It is therefore plausible that the shallow waters and high SST's may have resulted in increased evaporation and cloudiness especially during the warmer summer season, which combined with solar insolation minima and with high amounts of wind-blown snow during the winter, resulted in marginal glacier-friendly conditions on Mount Olympus during LG2.

Phase LG3 of glacier shrinking under cold and dry climate on Mount Olympus, finds firm evidence in Šara Range, Rila Mountains and Mount Uludağ (Figs 1 and 11). Similarly to LG1, glacial phase LG3 is characterized by a decoupling between solar insolation, which was increasing, and temperature and precipitation, which were low; these conditions were rather favourable to glacier shrinking (Fig. 12A). The cold and dry conditions during LG3 are related to the Younger Dryas, which affected the terrestrial and marine systems of the broader region to an equal or even to a greater extent than LG1. Terrestrial cooling in the SW Balkans (Lake Maliq) was 11°C and P_{ann} was reduced by 50% to 300-400mm (Bordon et al., 2009). The climatic conditions during LG3 resulted to minor forest contraction in the intramontane basin of Ioannina Lake (Lawson et al., 2004), located 35km south of Mount Tymphi and 135km west of Mount Olympus (Fig. 1), but did not favor the formation of glaciers in Mount Tymphi (Pope et al., 2015). This implies that during LG3 moisture availability in the SW Balkans was marginal to sustain forest vegetation

754 but not to promote glaciation in Mount Tymphi, the opposite holding truth for Mount Chelmos in southern
755 Greece (Fig. 1), where two samples with ^{36}Cl ages of 11.4 ± 1.4 ka and 14.2 ± 1.7 ka, were recalculated
756 using the same parameters as in this study (Pope et al., 2015, Fig. 11). The Younger Dryas glaciation in
757 Mount Chelmos has been attributed to a southern deflection of the Mediterranean storm tracks (Pope et al.,
758 2015), in a similar manner that occurred during the LGM (Kuhlemann et al., 2008). A different paleoclimatic
759 pattern in the SE Balkans arises from ELA reconstructions based on geomorphological evidence and
760 stratigraphic correlations from the Rila Mountains (Kuhlemann et al., 2013), and from SED's from Šara
761 Range (Kuhlemann et al., 2009), from Mount Uludağ West Ski area (Zahno et al., 2010) and from TZ and
762 MK cirques on Mount Olympus (Fig. 11). Glacier shrinking on Mount Olympus during LG3, was coupled
763 by a less pronounced drop in SST's, which was in the order of 6°C (Gogou et al., 2007, Kothoff et al.,
764 2011). A similar situation occurred in the north Aegean Sea, during the "8.2ka event", and was characterized
765 by considerably colder conditions, when continental cooling in Tenaghi Philippon and the Rhodope
766 Mountains (Fig. 1) was much more intense (Pross et al., 2009), in comparison to the north Aegean
767 borderlands (Kothoff et al., 2008, 2011). LG3 cold and dry climatic conditions had minor effects on the
768 vegetational record of Tenaghi Phillippon (Fig. 12E) and on the hydrological and sedimentological
769 processes of Lakes Prespa and Ohrid (Fig. 12C, D). More pronounced impacts during the LG3 glacial phase,
770 are recorded on the vegetation distribution (reduction of *Pinus* and *Quercus* pollen, Fig. 12G, H) and on the
771 sedimentation regime (Fig. 12 I, J decreasing fluvial inputs and coarsening trends of the sediments) of the
772 north Aegean Sea borderlands. This zonal partitioning of glacial behavior during LG3, is also evident in
773 marine and other terrestrial records with milder conditions to the Ionian and Adriatic Seas (the main source
774 of moisture for Mount Tymphi) and colder and dryer conditions in the Aegean and Marmara Seas (Kothoff
775 et al., 2011 and references therein).

776 The environmental conditions following LG3 during the Pleistocene–Holocene boundary recovered
777 gradually towards the establishment of warm and wet climate. This change is reflected in a number of
778 lacustrine and marine proxies among the few that are considered here (Fig. 12). These proxies include the
779 gradual reduction of steppic taxa from 11.5 ka, and increasing productivity in Lakes Prespa, Ohrid, Dorjan
780 (Aufgebauer et al., 2012, Vogel et al., 2010, Zhang et al., 2014) and in the Aegean Sea (Kothoff et al., 2008,
781 2011). A suite of marine proxies suggest lagged and stepwise increases in SST (Fig. 12F), precipitation, and
782 river runoff (Fig. 12I) with peak values at 9.6 ka (Fig. 12I), contemporary with the glacier standstill phase
783 HOL1 at 9.6 ± 1.2 ka (Fig. 10F, 11). The climatic thresholds responsible for HOL1 phase, are also
784 responsible for the slow recovery of *Quercus* forests (Fig. 12H), for increasing fluvial inputs (Fig. 12I) and
785 for a meltwater pulse in the north Aegean Sea (Fig. 12B) that reduced the surface salinities (Kothoff et al.,
786 2008). In addition to the increasing fluvial runoff, this meltwater pulse was likely fostered through the
787 preservice of the glacial ice and extensive permanent snowfields in the northeast Mediterranean mountains
788 that resulted to subsequent discharge of both surface and karstic runoff during the melt out season, in
789 agreement with Ehrmann et al. (2007). Once these climatic thresholds were surpassed at ~ 9.6 ka, complete
790 deglaciation of the MK cirque occurred during the onset of Sapropel 1 formation (Kothoff et al., 2008).
791 Despite the fact that the mid-Holocene (9.5 – 3.0 ka) period has been characterized by considerable
792 millennial and centennial-scale variability in terrestrial and marine systems of northeast Mediterranean (e.g.
793 Rohling et al., 2002, Ehrmann et al., 2007, Kothoff et al., 2008, Tryantaphyllou et al., 2009, Schmiedl et al.,
794 2010, Styllas and Ghilardi, 2017), no evidence of glacier fluctuations are represented in the ^{36}Cl chronology
795 of MK cirque. These findings does thus not confirm the hypothesis postulated in our preceding work that the
796 MK glaciers advanced during the mid-Holocene (Styllas et al., 2016). It is likely that the position of sea-
797 level nearly at its present location (18km from the TZ and MK cirques) together with peak solar insolation,
798 and high SST's and summer (mean July) air temperatures, are among the key factors that prohibited the
799 return to glacier friendly conditions of sufficient duration to create glacial landforms in MK cirque. It has to
800 be acknowledged that higher mean July air temperatures by $1\text{--}2^\circ\text{C}$ (Samartin et al., 2017), compared to our
801 previously considered pollen-derived summer temperatures (gridded data by Mauri et al., 2015), may have
802 also contributed to a "glacier free" mid Holocene on Mount Olympus. But, in the case that there had been
803 glacier advance(s) during the mid-Holocene, they may be overridden by the subsequent late Holocene
804 glacier expansions. The reduction of solar insolation during the Late Holocene (Fig. 12A), along with wet
805 conditions between 3.3 and 2.6 ka recorded in the speleothem record of Skala Marion (Fig. 1), as well as in
806 other eastern Mediterranean speleothem records (Psomiadis et al., 2018), are likely the main driving
807 mechanisms of glacier expansion and stabilization of moraine MK-C3 at 2.5 ± 0.3 ka (Fig. 7) during HOL2
808 glacial phase (Figs 11 and 12). HOL2 is concomitant with a major soil erosion period recorded in the

809 geochemical and isotopic record of Lake Dorjan (Fig. 1), (Rothacker et al., 2018); the authors ascribe this
810 event to human activity, but given the wet and cold conditions that resulted to a glaciation phase, a
811 combination of climatic and anthropogenic factors cannot be ruled out. HOL2 likely initiated around 2.8 ka
812 during a major solar insolation low, the Homeric Minimum (Wirth and Sessions, 2016) and its duration
813 matches the Bond 2 event (Bond et al., 2001), which like other Bond events during the Holocene, has been
814 found to correspond to periods of high hydrologic activity of the Aliakmon River (Fig. 1) (Styllas and
815 Ghilardi, 2017). Therefore, it is plausible that the climatic and environmental conditions during HOL2, with
816 a glacier expanse in Megala Kazania cirque and a local ELA of 2250m, were associated with the existence
817 of permanent snowfields in the Throne of Zeus cirque and other locations above the local ELA. This might
818 have fed the imaginations of the Ancient Greeks and may have given Mount Olympus its name (“*Ὀλύμπος*”),
819 which means the “ever shining mountain” (Curtius, 1879). Homer himself refers to Mount
820 Olympus permanent snows in many occasions in both ancient Greek poems of Iliad and Odyssey (Nezis,
821 2000). From a paleoclimatic standpoint, glacial phase HOL2 exhibits special interest, as it can provide
822 quantitative estimates on annual precipitation and temperature during a period when human activities were
823 intense enough to alter the pollen composition of lacustrine and marine records, obscuring the paleoclimatic
824 information (e.g. Kothhoff et al., 2008, Vogel et al., 2010, Kothhoff et al., 2011, Aufgebauer et al., 2012,
825 Francke et al., 2013, Zhang et al., 2014, Pross et al., 2015, Rothacker et al., 2018, Wulf et al., 2018).
826 The last phase of glacial activity depicted from our SED chronology (HOL3) is derived from blocky
827 moraine MK-C4, which is located 100m below the present-day terminus of Mount Olympus extant
828 snowfields (Fig. 5). HOL3 occurred at 0.64 ± 0.08 ka (AD 1320 – 1480) during the latter part of the
829 Medieval Climate Anomaly (MCA) and the early part of the Little Ice Age (LIA), a transition that in Eastern
830 Mediterranean was characterized by a shift from wet to dry conditions (Roberts et al., 2012). Even though
831 HOL-3 is constrained by only one SED, tree-ring proxy data from Mount Olympus, point to warmer than
832 present summer (June – September) temperatures between AD 1500 – 1700 (Klesse et al., 2014) and this
833 trend is also reflected in the north Aegean Sea SST record (Gogou et al., 2016). Higher than present summer
834 temperatures and increased SST’s, on Mount Olympus and the north Aegean Sea, were coupled by regional
835 aridity as manifested by the decrease of allochthonous sediments in Lake Dorjan (Francke et al., 2013)
836 around AD 1550 and the combination of these conditions most likely triggered the early LIA glacier retreat
837 of Megala Kazania paleoglacier. A reverse signal is observed in the western Balkans, as the interval between
838 AD 1500 – 1800, is characterized by wet climatic conditions (Morellón et al., 2016, Koutsodendris et al.,
839 2017). HOL3 and LG3 glacial phases, emphasize the existence of an E-W climatic partitioning along the
840 southern Balkan Peninsular, evident in the Mediterranean basin during the last millennium (Roberts et al.,
841 2012), but the boundary of this see-saw climatic pattern needs to be better defined by additional proxy-based
842 studies (Koutsodendris et al., 2017), including the glacial record.

843 6. Conclusions

845 In this study, we propose a new Late-glacial and Holocene chronology of glacial phases of Mount Olympus
846 and compare it with existing SED based glacier chronologies from the northeast Mediterranean mountains.
847 Our glacial chronology is based on *in situ*-produced ^{36}Cl -based cosmic ray exposure dating of 20 glacially
848 transported boulder and bedrock samples from the east facing Throne of Zeus (TZ) and northwest oriented
849 Megala Kazania (MK) cirques, which are consistent with the stratigraphic positions of the
850 geomorphologically distinct frontal, push, hummocky and blocky moraines. The new glacial chronology
851 from Mount Olympus is complementary, but greatly refines the existing SED-based glacier chronologies
852 from Mount Pelister, Galicica Mountains, Šara Range, Rila Mountains and Mount Uludağ.
853 The first Late-glacial phase of glacial variability (LG1) is characterized by moraine stabilization at ~15.5 ka
854 under overall dry and cold conditions, followed by deglaciation due to increasing temperatures and
855 precipitation during its later stage at ~14.0 ka. Solar insolation maxima and low precipitation during LG1,
856 likely restricted the glacier expansions within the higher cirques and at elevations between 2000 and 2200m.
857 The climatic conditions that followed LG1, are characterized by considerable increases in precipitation and
858 temperature, but in the highest cirques of Mount Olympus such increases were less pronounced and forced a
859 shift to marginally glacial conditions with intense aeolian activity from western directions and solar
860 insolation minima at ~13.5 ka during LG2. A sharp drop in precipitation, air temperatures and north Aegean
861 SST’s marked the beginning of phase LG3 with immediate impacts on the high cirques of the northeast
862 Mediterranean mountains, which experienced a return to glacial conditions at ~12.5 ka. On Mount Olympus,

863 LG3 was expressed by a significant drop in air temperatures and overall glacier shrinking, whereas milder
864 and/or drier conditions occurred to the west of the Pindos Mountains (southwestern Balkans). Within error
865 uncertainties of our new SED and mean landform ages, glacial phases LG1-3 can be tentatively correlated to
866 the Older Dryas, Bölling/Alleröd and Younger Dryas periods of the Greenland isotope record.
867 The Pleistocene / Holocene transition saw a gradual and stepwise return to milder conditions that are
868 recorded in a large array of lacustrine and marine proxies. Above 2000m, the deglaciation followed the
869 reorganization of the atmosphere, as high precipitation amounts and cooler summer temperatures likely
870 forced a glacier standstill or re-advance phase (HOL1) at ~9.6 ka, followed by the onset of very humid and
871 warm conditions in the north Aegean Sea. The mid-Holocene glacial evolution was likely characterized by
872 complete deglaciation, in phase with high summer temperatures and solar insolation. A decline in solar
873 insolation combined with climate deteriorations associated with North Atlantic's Bond Events, resulted in a
874 glacier expansion in Megala Kazania cirque during the late Holocene (HOL2). The initiation of phase HOL2
875 might have started around 2.8 ka during the Homeric solar low, combined with wet conditions and
876 subsequent moraine stabilization occurred at ~2.5 ka. The last phase of glacial activity on Mount Olympus
877 as well as on other northeast Mediterranean mountains (HOL3) was less extensive than HOL2 and was
878 restricted close to the cirque headwalls. Phase HOL3 corresponds to the end of the Medieval Climate
879 Anomaly and the beginning Little Ice Age, as evidenced from one boulder dated to ~0.6 ka.
880 In summary, we present here for the first time a glacial chronology of the northeast Mediterranean
881 mountains that spans the Late Glacial and the Holocene and is consistent with a number of terrestrial and
882 marine proxies. An *out-of-phase* behaviour between the a solar insolation peak and glacial phases is
883 observed during the Late-glacial, but this pattern gets back into phase during the Holocene, as glacial phases
884 occur during solar insolation lows. We show that in addition to the early Little Ice Age glacier advance, two
885 Holocene glacial phases are recorded in the glacial geomorphology on Mount Olympus, during the early and
886 late Holocene. In comparison with the other cirques under consideration, these Holocene glacial phases were
887 more pronounced in Megala Kazania cirque due to its topographic characteristics and proximity to the north
888 Aegean Sea.

891 **Acknowledgements**

892
893 This work was financed by the Scientific Studies Program of John S. Latsis Foundation, Athens, Greece.
894 The French AMS national facility ASTER (CEREGE, Aix en Provence) is supported by the INSU/CNRS,
895 the ANR through the "Projets thématiques d'excellence" program for the "Equipements d'excellence"
896 ASTER-CEREGE action and IRD. Michael Styllas is particularly thankful to Basil Davis (Université de
897 Lausanne) and to Stefan Klesse (University of Arizona) for their fruitful discussions, commends and data
898 sharing during the course of the project "Did the Ancient Greek Gods Ever Got Cold? Examining the
899 Holocene Glacial History of Mount Olympus, Greece". We are thankful to two anonymous reviewers for
900 their constructive comments that further improved the manuscript.

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