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1 **Glaciations on ophiolite terrain in the North Pindus mountains, Greece:**  
2 **new geomorphological insights and preliminary <sup>36</sup>Cl exposure dating**

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13 **Abstract**

14 *A glacial geomorphological analysis of three valleys on Mt Mavrovouni (North Pindus*  
15 *Mountains, Greece) is presented alongside a pilot study using cosmogenic <sup>36</sup>Cl to obtain*  
16 *surface exposure ages from iron-rich ophiolite glacial and periglacial boulders. At least*  
17 *three distinct morphostratigraphic units of glacial (moraines) and periglacial (relict pronival*  
18 *ramparts) origin have been identified. Four <sup>36</sup>Cl surface exposure ages were obtained from*  
19 *the stratigraphically youngest glacial and periglacial deposits. Although this limited dataset*  
20 *with relatively large uncertainties cannot support a robust geochronology, the ages are*  
21 *consistent with the <sup>36</sup>Cl-based chronologies of limestone-derived moraines on Mt Tymphi*  
22 *(NW Greece) and Mt Chelmos (S Greece), confirming that the last glaciers on this massif*  
23 *formed during the Last Glacial Maximum as also indicated by other studies in the Pindus*  
24 *mountains. At the same time it provides confidence in the suitability of <sup>36</sup>Cl dating for iron-*  
25 *rich samples, such as ophiolites, using an updated <sup>36</sup>Cl model that incorporates improved*  
26 *production rates for iron spallation. The presented preliminary chronology of moraines and*

27 *pronival ramparts is based on those ages as well as on local and regional*  
28 *morphostratigraphic correlations. The stabilisation of the most extensive Late Pleistocene*  
29 *glaciers took place during the Last Glacial Maximum, at  $27.0 \pm 6.5$  ka whereas the presence*  
30 *of pronival ramparts dated at  $20.2 \pm 4.8$  ka suggests persisting cold and arid conditions.*  
31 *Older, still undated glacial deposits exist lower in the valleys which can be attributed to the*  
32 *Middle Pleistocene major glaciation phases (MIS 12 / MIS 6), based on their relative*  
33 *morphostratigraphic position within the glacial sedimentary sequence.*

34 **Key words:** glacial geomorphology; Greece; ophiolites; cosmogenic  $^{36}\text{Cl}$  exposure age  
35 dating, Last Glacial Maximum

## 36 **1 Introduction**

37 The former extents of mid-latitude mountain glaciers represent a sensitive record of past  
38 changes in climate as their relatively small size means that they rapidly change volume in  
39 response to mass balance changes caused by regional or local climatic changes (Bahr et al.,  
40 1998; Oerlemans, 2005; Woodward, 2009; Vogiatzakis, 2012). Mapping and dating of past  
41 ice margins, documented by glacial sediments and landforms, not only provides insights into  
42 past climate change but also provides valuable information for understanding landscape  
43 evolution (Owen et al., 2002; Svendsen et al., 2004).

44 In order to compare palaeoglaciers in different geographical settings, a useful tool is the  
45 Equilibrium Line Altitude (ELA), i.e. the elevation of a glacier where annual accumulation  
46 and ablation are equal, which is governed by air temperature and precipitation (Ohmura and  
47 Boettcher, 2018). Moreover, as temperatures at a regional level may practically depend on  
48 elevation rather than latitude (e.g. Greece, central Italy; Katsoulakos and Kaliampakos, 2014;  
49 Papada and Kaliampakos, 2016), ELAs in this case can be considered to be mainly governed  
50 by precipitation and therefore constitute a direct precipitation proxy.

51 Evidence of former glaciation is abundant throughout the mountains of Greece (e.g.  
52 Niculescu, 1915; Mistardis, 1935; Hughes et al., 2006a, 2006b; Styllas, 2018; Leontaritis et  
53 al., 2020 and references therein). The Pindus Mountains in north-western Greece is  
54 historically a key-region for the study of the glacial history of the Balkans and the  
55 Mediterranean (Woodward and Hughes, 2011). The first pioneer geographical studies  
56 reporting glacial features took place early in the 20th century (Niculescu, 1915; Bourcart,  
57 1922; Louis, 1926; Sestini, 1933; Mistardis, 1935) whereas systematic geomorphological  
58 mapping was initiated decades later (Palmentola et al., 1990; Boenzi et al., 1992). Hughes et  
59 al. (2006a) used U-series dating on the most extensive moraines on Mt Tymphi to show that  
60 the most intense glacial phases correspond to MIS 6 (Vlasian Stage: ca. 190-130 ka) and  
61 further back to MIS 12 (Skamnellian Stage: ca. 480-430 ka) during the Middle Pleistocene.  
62 The stratigraphically younger Late Pleistocene moraines on Mt Tymphi were only recently  
63 dated to the Last Glacial Maximum (LGM: 27.5-23.3 ka; Hughes and Gibbard, 2015) with  
64  $^{36}\text{Cl}$  surface exposure dating on limestones (Allard et al., 2020). Surface exposure ages from  
65 Mt Chelmos in southern Greece ( $^{36}\text{Cl}$  ages from limestones ; Figure 1) indicate that this  
66 glacial phase is likely to correspond to an interval c. 5-10 ka prior to the global LGM (Pope et  
67 al., 2017). Today, the Late-glacial (17.5-11.7 ka) features of the region remain largely  
68 neglected because their presence is limited to the ophiolitic Mt Smolikas (Leontaritis et al.,  
69 2020; Figure 1), with its peculiar lithology preventing the dating of its moraines with absolute  
70 dating methods (Woodward and Hughes, 2011). Thus, the response of glaciers in the Pindus  
71 Mountains to the cold events of this interval remains poorly elucidated. A lingering question  
72 is if this evidence supports a local (intra-Pindus) and regional (across Greece) chronological  
73 and morphostratigraphic framework to describe Quaternary glaciations or if local topographic  
74 and climatic factors dominated glacial growth during specific glacial phases like for example  
75 Holocene glaciations in Montenegro (Hughes, 2007; 2010a).

76 Terrestrial cosmogenic nuclides have been used to date glacial landforms all over the  
77 world (e.g. Gosse et al., 1995; Finkel et al., 2003; Balco and Schafer, 2006; D'Arcy et al., 2019;  
78 Anjar et al., 2021), but the ophiolitic lithology in the Pindus restricts possible nuclides to only  
79  $^3\text{He}$  and  $^{36}\text{Cl}$  (Leontaritis, 2021), so  $^{36}\text{Cl}$  was chosen due to the ability to handle any possible  
80 variability in the lithology (e.g. lack of sufficient olivines; see Ivy-Ochs and Kober, 2008).  
81 Although  $^{36}\text{Cl}$  exposure dating has been successfully applied on serpentized harzburgites to  
82 date glacial deposits (see Sarikaya et al., 2008), it has not yet been fully characterised. This is  
83 the only option for carbonate rocks as well and so has been used to date the above-mentioned  
84 moraines in the limestone massifs of Mt Tymphi (Allard et al., 2020) and Mt Chelmos (Pope  
85 et al., 2007). However, carbonate rocks are also more susceptible than others to the effects of  
86 surface erosion, so this potentially increases the uncertainty associated with those ages  
87 (Allard et al., 2020, Sarikaya et al., 2020).

88 The aim of this paper is to establish a preliminary chronology of the glacial deposits on  
89 Mt Mavrovouni in the Pindus Mountains based on a detailed glacial geomorphological study  
90 and the preliminary timing constraint of its ophiolite-dominated glacial deposits through a  
91 pilot study using cosmogenic  $^{36}\text{Cl}$  surface exposure dating. The main objective is to confirm  
92 that the formation of the stratigraphically youngest glacial deposits in this massif falls within  
93 the LGM as suggested by morphostratigraphic correlations of moraines within the Pindus  
94 mountains (Mt Tymphi, Mt Smolikas and Mt Mavrovouni; Leontaritis, 2021; Figure 1).  
95 Thus, in this pilot study, sampling was limited to these sediments, initially correlated with  
96 Late Pleistocene glacial activity. The small number of ages allows only for a preliminary  
97 timing constraint of the respective cold events on Mt Mavrovouni and additional data will  
98 still be needed to establish a tighter and more reliable chronology. However, the presented  
99 chronology is sufficient to confirm the viability of cosmogenic  $^{36}\text{Cl}$  surface exposure dating  
100 using new production rates for Fe spallation on iron-rich ophiolites. Given the fact that some

101 of the highest peaks of the Pindus and the wider Balkans are formed in this lithology, it sets  
102 the agenda for future research on similar glaciated terrains.

## 103 **2 Regional Setting**

104 Mt Mavrovouni is located in the north Pindus Mountains in NW Greece (Figure 1), only  
105 20km to the east/southeast of Mt Tymphi (2497m a.s.l.) and 30km to the southeast of Mt  
106 Smolikas (2637m a.s.l.). Its highest peak (Flega) reaches an altitude of 2157m a.s.l.;  
107 considerably lower than the neighbouring mountains. Mt Mavrovouni, alongside, Mt  
108 Smolikas and Mt Vasilitsa (2248 m a.s.l.) belongs to the Northern Pindus ophiolitic complex,  
109 part of a nappe which is tectonically overthrust onto the Eocene flysch of the Pindus Zone  
110 (Dupuy et al. 1984). In particular, Mt Mavrovouni belongs to the Dramala Complex subunit  
111 which represents oceanic mantle and part of its crustal sequence (Pelletier et al., 2008). It  
112 primarily consists of slightly to highly serpentized harzburgites of Mid-Jurassic age while  
113 subordinate lithologies include dunite, pyroxenite and a range of ultramafic cumulate rocks  
114 (Jones and Robertson, 1991; Robertson, 2002; Pelletier et al., 2008). Mesozoic limestones  
115 and Palaeozoic crystalline schists are also present locally, at the overthrust nappe of the  
116 ophiolitic complex (IGME, 1959).

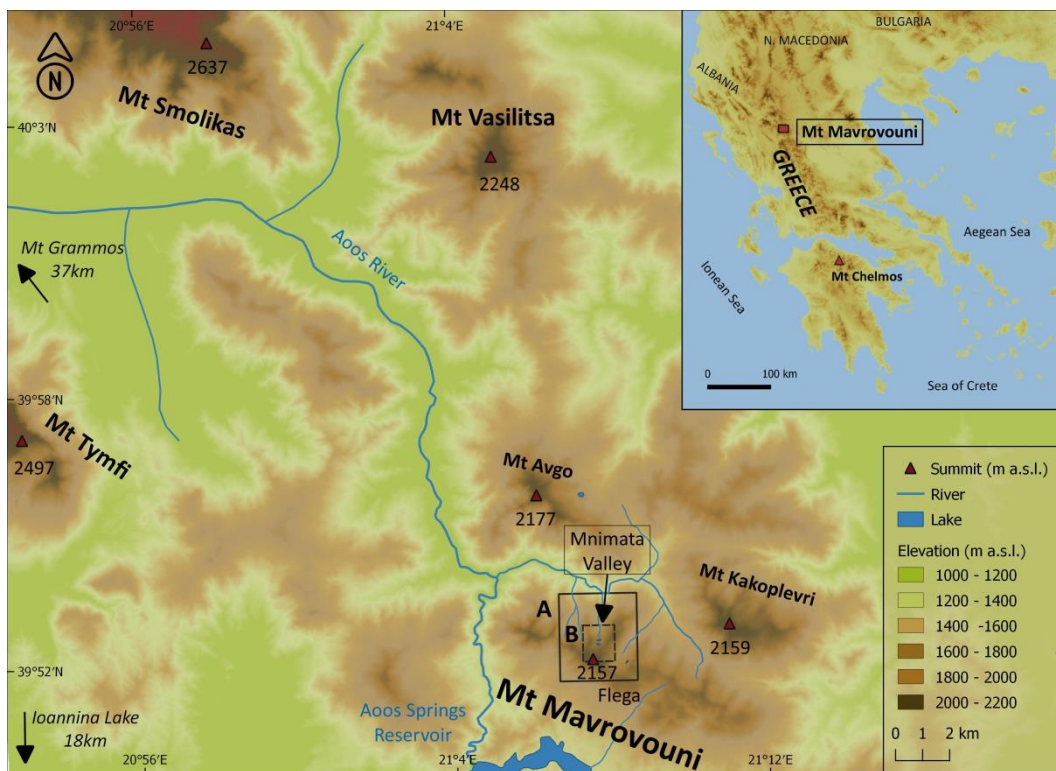
117 In common with its ophiolitic neighbours, Mt Mavrovouni presents a less complex  
118 physiography compared with the limestone dominated Mt Tymphi, as it is a generally  
119 rounded mountain with many steep V-shaped valleys and only a few and relatively short  
120 cliffs. Evidence of glaciation is present on its northern slopes. The glacial sedimentary  
121 morphosequence (i.e. the spatial distribution of moraines along a valley indicating phases of  
122 glacial advance/standstill) is fragmented with the exception of the U-shaped Mnimata valley  
123 (Figure 2). Focus was given on constraining the timing of processes that formed the highest  
124 and therefore stratigraphically youngest glacial/periglacial deposits in the Mnimata valley,



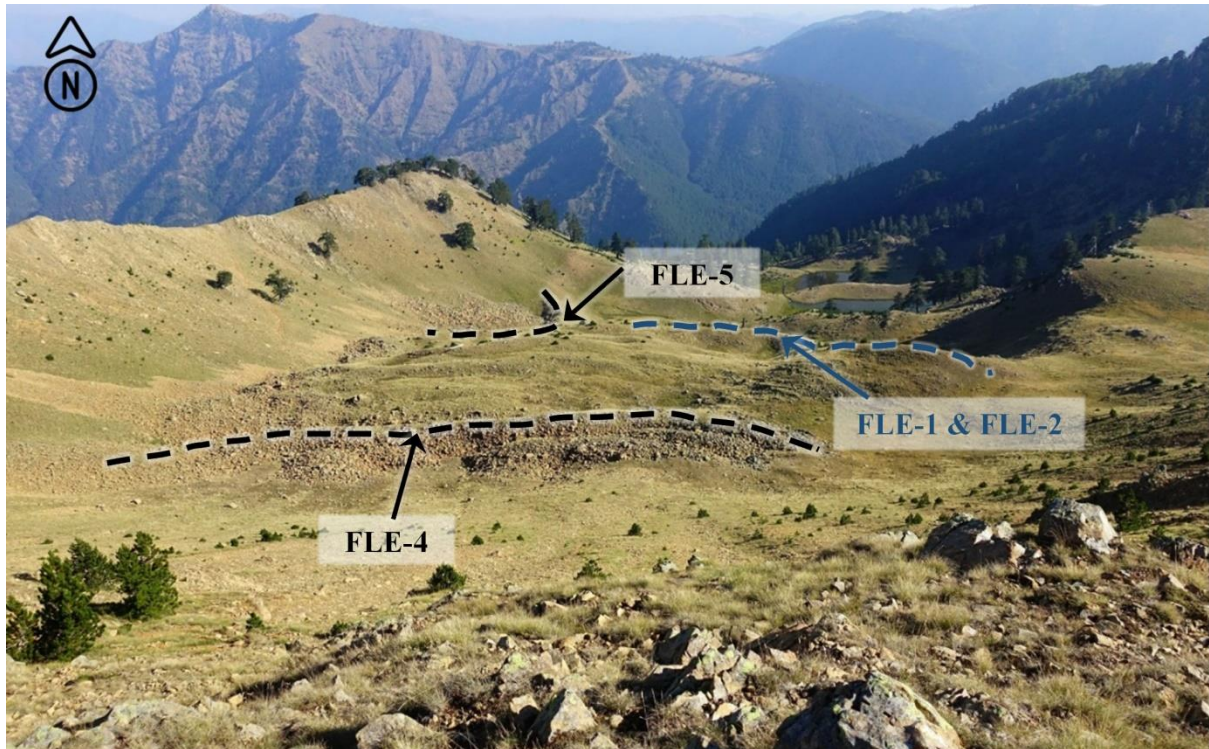
125 which, according to evidence elsewhere in Greece, are expected to be Late Pleistocene in age  
126 (Leontaritis et al., 2020 and references therein).

127 No glaciers, permafrost or permanent snow fields are present today. Snow cover on the  
128 northern slopes above 1800m a.s.l. lasts from early December to late May. Direct  
129 climatological data are available from four high-altitude local stations at distances <20km  
130 (Vovoussa - 1000m a.s.l., Greveniti - 976m a.s.l., Elatochori - 1014m a.s.l. and Metsovo -  
131 1160m a.s.l.; precipitation data from Fotiadi et al., 1999; temperature data from Gouvas and  
132 Sakellariou, 2011; reference period >20yr).. Extrapolating temperature data to a  
133 representative altitude of 2000m a.s.l. with a lapse rate of 0.6°C per 100m of altitude  
134 (Leontaritis, 2021) estimated the mean annual and monthly (July, January) temperatures at  
135 4.5°C, 13.8°C and -3.0°C respectively. Precipitation data were extrapolated using a  
136 precipitation gradient of 55 mm per 100 m altitude for this area (Hughes, 2004) and the mean  
137 annual precipitation at 2000m a.s.l. is expected to exceed 2000mm (Leontaritis, 2021)  
138 confirming earlier suggestions by Furlan (1977).

139



140 *Figure 1. Map of the study area on Mt Mavrovouni and location of the massif in the southern tip of the*  
141 *Balkan Peninsula (up-right inset map). Detailed glacial geomorphological maps of the areas denoted*  
142 *by boxes A and B are given in Figure 4 and Figure 6, respectively.*



143

144 *Figure 2. The Mnimata U-Shaped valley and the glacial/periglacial sedimentary sequence. Sample*  
145 *locations on terminal moraine (dashed blue line) and pronival rampart ridges (dashed black lines) are*  
146 *denoted. Photo looking north from the Flega Peak at 2157 m a.s.l. (Photo: August 2016).*

### 147 **3 Methods**

#### 148 **3.1 Geomorphological mapping**

149 Geomorphological mapping was conducted following the principles of the idealised  
150 approach proposed by Chandler et al. (2018) for glacial geomorphological mapping in alpine  
151 environment which involves remotely-sensed data and field mapping. Accordingly,  
152 reconnaissance analysis was performed on available topographic maps (e.g. Anavasi, 2016;  
153 Open Street Maps), the regional 1:50,000 Geological Sheets of the Greek Institute for  
154 Geological and Mineral Exploration (IGME, 1959), and high-resolution satellite imagery



155 (Google Earth, Bing Maps). The Digital Elevation Model was provided by the National  
156 Cadastre at a spatial resolution of 5m. In the field, the reconnaissance sketch-map was refined  
157 and glacial, glaciofluvial and periglacial landforms and sediment deposits were mapped and  
158 recorded using a commercial GPS/Glonass device (Garmin 64s) over three fieldtrips in the  
159 summers of 2016, 2017 and 2018. Additionally, UAV-captured (Unoccupied Aerial Vehicle)  
160 aerial photos and videos were taken from the upper Mnimata valley. The digital mapping and  
161 the creation of the glacial geomorphological maps were conducted in the open source QGIS  
162 environment.

### 163 **3.2 Morphostratigraphic classification of glacial and periglacial features**

164 In this paper, we used morphostratigraphic classification of the different glacial and  
165 periglacial features following the principles in Hughes (2010b). Relict pronival ramparts  
166 (periglacial or transitional units) were distinguished from moraines using multiple  
167 identification criteria (see Hedding and Sumner, 2013; Matthews et al., 2017 and references  
168 therein). Units were separated by morphostratigraphic criteria such as relative stratigraphic  
169 position, elevation, sediments composition (e.g. clasts size and shape, presence of fines),  
170 evidence of exhumation and soil development. Local (i.e. within the same massif) and  
171 regional (e.g. within NW Greece) morphostratigraphic correlations allowed the utilization of  
172 spatial relationships between individual glacial/periglacial landforms in order to extrapolate  
173 existing chronologies of the Pleistocene glacial sequence. Due to the limited availability of  
174 dateable sites or material this approach has been successfully applied in glacial  
175 geomorphology worldwide, underlying many glacier reconstructions (e.g. Lowe and Walker,  
176 1997; Chadwick et al., 1997; Richards et al., 2000; Benn and Ballantyne, 2005; Lukas, 2006).

### 177 **3.3 Cosmogenic $^{36}\text{Cl}$ surface exposure dating**

178 The ophiolitic lithology of the Northern Pindus Mountains (Mt Mavrovouni, Mt  
179 Smolikas, Mt Vasilitsa and Mt Grammos; Figure 1) has so far prevented the age control of

180 glacial deposits with absolute dating methods (Hughes et al., 2006a; Woodward and Hughes,  
181 2011). In particular, widely used techniques for dating glacial deposits such as cosmogenic  
182  $^{10}\text{Be}$  or  $^{26}\text{Al}$  surface exposure dating, luminescence and U-series are unsuitable for such  
183 lithologies due to the complete lack of quartz, feldspar or secondary calcites (Fuchs and  
184 Owen, 2008; Ivy-Ochs and Kober, 2008). In theory, cosmogenic  $^{36}\text{Cl}$  surface exposure dating  
185 can be used for any rock type under most conditions (Ivy-Ochs and Kober, 2008) and it has  
186 successfully been applied to date glacial deposits on ophiolites and specifically on  
187 serpentinized harzburgites on Mt Sandiras, SW Turkey (Sarıkaya et al., 2008). In such  
188 lithologies, the relatively high concentrations of iron (Fe) and native Cl means that the main  
189 production of cosmogenic  $^{36}\text{Cl}$  is expected from neutron capture by  $^{35}\text{Cl}$  and spallation  
190 reactions in Fe (Gosse and Phillips, 2001; Marrero et al., 2016a). Because spallation from Fe  
191 usually comprises <2% of total production in most rocks (Marrero et al., 2016a), production  
192 from this pathway was not explicitly calibrated in earlier large-scale  $^{36}\text{Cl}$  calibration efforts  
193 (Phillips et al., 2001; Swanson and Caffee, 2001; Schimmelpfennig et al., 2011; Marrero et  
194 al., 2016a), leading to a potential for larger uncertainties in high-Fe samples (Gosse and  
195 Phillips, 2001; Asimov and Ivy-Ochs, 2009; Marrero et al., 2016a; Moore and Granger,  
196 2019). This study incorporates new Fe-spallation production rates from Moore and Granger  
197 (2019) in order to reduce uncertainty on high-Fe sample ages (Anjar et al., 2021).

### 198 **3.3.1 Sample collection and characterization**

199 The sampling strategy for  $^{36}\text{Cl}$  cosmogenic nuclide dating was influenced by the objective  
200 of this study which was to confirm that the last glaciation on Mt Mavrovouni formed during  
201 the last glacial cycle as well as by the limited resources available for  $^{36}\text{Cl}$  analyses. Therefore,  
202 this study focused from a geochronological point of view on the stratigraphically youngest  
203 glacial/periglacial deposits on Mt Mavrovouni according to the geomorphological analysis.  
204 More specifically, sampling was aimed at the sequence of glacial/periglacial deposits in the

205 upper Mnimata valley (Figure 2). Four samples were collected following the principles in  
206 Gosse and Phillips (2001). Two samples (FLE-1 and FLE -2) were collected from the ridge of  
207 a terminal moraine at the lowermost part of this sequence whereas two more samples were  
208 collected from the ridges of two relict ramparts immediately below the cirque walls (FLE-4  
209 and FLE-5). The approach to date the stratigraphically youngest and oldest deposits in this  
210 sequence was followed in an effort to best exploit the small number of available analyses in  
211 order to constrain the timing of this cold stage.

212 All sampled boulders are of ophiolitic lithology. Samples were obtained with the use of a  
213 hammer and chisel from the upper 2-3cm of a flat rock surface as far away as possible from  
214 the boulder's edges to minimise edge effects (Gosse and Phillips, 2001; Masarik and Wieler,  
215 2003; Darvill, 2013). Topographic shielding for the four samples was determined with the  
216 use of an open-access DEM-based GIS toolbox dedicated to the calculation of cosmic  
217 radiation topographic shielding for discrete sample points on the available 5m-resolution  
218 DEM (Li, 2018; available online at <https://web.utk.edu/~yli32/programs.html>).

219 The highly-serpentinized ophiolitic rocks show very limited signs of weathering implying  
220 very low surface erosion rates as also indicated by ice-polished bedrock with well-preserved  
221 striations below the lower Flega tarn (Figure 3). A surface erosion correction factor for the  
222 calculation of the final ages was thus not considered. Also, information about contemporary  
223 and Late Pleistocene snow cover is not available so this was not considered. Styllas et al.  
224 (2018) showed that on Mt Olympus even when quite a significant snow correction factor was  
225 considered (0.94), the resulting deviations on calculated Late-glacial exposure ages were in  
226 the scale of a few hundred years. Taking into consideration the calculated age uncertainties,  
227 such deviations are practically negligible and the calculated (non-corrected and thus  
228 minimum) ages can be associated to the age of moraine deposition. The location and  
229 characteristics of the collected samples are summarised in Table 1.



231  
 232 *Figure 3. Ice-polished bedrock with well-preserved striations below the lower Flega tarn is indicative*  
 233 *of very low surface erosion rates. This outcrop of ice-polished bedrock was exposed after the glacier's*  
 234 *retreat and as indicated by field evidence the preservation of the striations should not be attributed to*  
 235 *soil cover protection: soil in this upper part of the massif has been developed mostly on depressions*  
 236 *whereas rocky exertions generally lack any soil cover (Photo August 2016).*

237 *Table 1. Location and characteristics of collected  $^{36}\text{Cl}$  samples.*

Sample ID	Latitude (N°)	Longitude (E°)	Elevation (m a.s.l.)	Sample thickness (cm)	Sample Density (g cm <sup>-3</sup> )	Topographic Shielding Factor (-)
FLE-1	39.8737	21.1204	2017	2.0	2.7	0.978
FLE-2	39.8735	21.1205	2017	2.5	2.7	0.985
FLE-4	39.8725	21.1190	2038	3.0	2.7	0.978

FLE-5	39.8749	21.1193	1978	3.0	2.7	0.989
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### 238 3.3.2 Samples preparation AMS measurements

239 Rock samples were crushed and sieved at the National Technical University of Athens.  
 240 Subsequently, the 250-500  $\mu\text{m}$  fraction underwent chemical processing and measurement at  
 241 the Purdue Rare Isotope Measurement Laboratory (PRIMElab) at Purdue University.  
 242 Samples were leached in 5% dilute nitric acid ( $\text{HNO}_3$ ) several times, had  $\sim 1$  mg of chlorine  
 243 carrier added (Table 2), dissolved with concentrated hydrofluoric (HF) and nitric acids,  
 244 heated in a  $60^\circ\text{C}$  water bath, then fluoride by-products were removed by centrifugation. Silver  
 245 chloride ( $\text{AgCl}$ ) was precipitated from samples, then it was purified using barium nitrate  
 246 [ $\text{Ba}(\text{NO}_3)_2$ ] and anion columns before being dried and pressed into cathodes.

247 *Table 2. AMS analytical data for  $^{36}\text{Cl}$  samples. Standard Deviation given in parentheses. \*Carrier:*  
 248  *$^{35}\text{Cl}/^{37}\text{Cl}$  ratio 273. Total Cl content was determined using standard isotope dilution calculations*  
 249 *(Desilets et al., 2006).*

Sample ID	Mass (g)	Carrier* added (mg Cl)	$^{36}\text{Cl}/\text{Cl}$ ratio ( $\times 10^{-15}$ )	$^{35}\text{Cl}/^{37}\text{Cl}$ stable ratio	Cl content (ppm)	Measured $^{36}\text{Cl}$ (atoms $\cdot \text{g}^{-1}$ )
FLE-1	30.242	1.0805	108.5 (4.2)	5.48 (0.03)	58.89 (0.76)	177,500 (7,100)
FLE-2	30.188	1.0864	186.7 (5.8)	5.51 (0.06)	58.57 (1.62)	307,700 (10,900)
FLE-4	30.213	1.0841	161.7 (5.1)	3.90 (0.03)	188.59 (6.67)	622,800 (27,000)
FLE-5	30.221	1.0676	368.0 (8.4)	8.50 (0.02)	23.43 (0.22)	386,200 (8,900)

250 The  $^{36}\text{Cl}$  to total stable chlorine ratio ( $^{36}\text{Cl}/\text{Cl}$ ) and the stable  $^{35}\text{Cl}/^{37}\text{Cl}$  ratio were  
 251 measured by Accelerator Mass Spectrometry (AMS) on the 8 MV tandem accelerator at  
 252 PRIME Lab, normalising  $^{36}\text{Cl}/\text{Cl}$  measurements to NIST reference material SRM 4943  
 253 standards (Sharma et al., 1990). The AMS data for the samples alongside the calculated total  
 254 Cl content and  $^{36}\text{Cl}$  concentration for each sample are summarised in Table 2.

### 255 3.3.3 Cosmogenic $^{36}\text{Cl}$ exposure age calculations



256 The exposure ages of the samples were calculated using the CRONUScalc <sup>36</sup>Cl Exposure  
257 Age Calculator v2.2 (original code available in Bitbucket repository: v2.2,  
258 <https://bitbucket.org/cronusearth/cronus-calc/>; Marrero et al., 2016b). The time-dependent  
259 and nuclide-dependent Lifton-Sato-Dunai scaling (LSDn) scaling framework (Lifton et al.,  
260 2014) was selected, due to its fit to calibration data (Phillips et al., 2016). This model is based  
261 on equations from a nuclear physics model and incorporates dipole and non-dipole magnetic  
262 field fluctuations and solar modulation and includes nuclide-specific scaling factors (Lifton et  
263 al., 2014; Marrero et al., 2016b). Production rates used in the calculator are from Marrero et  
264 al. (2016a; 2021), with newly incorporated <sup>36</sup>Cl production rates from iron spallation from  
265 Moore and Granger (2019), as described in Anjar et al. (2021).

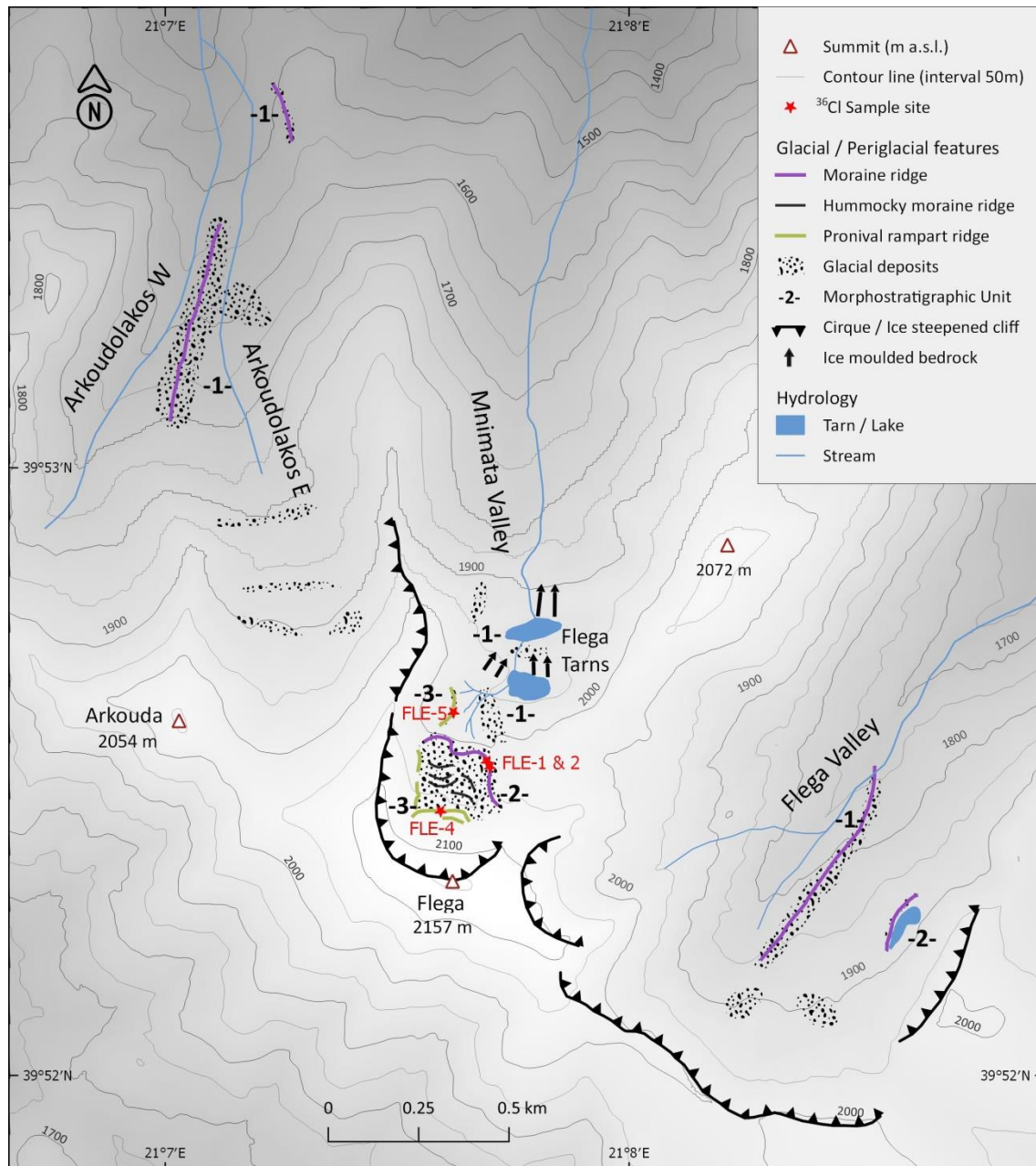
## 266 **4 Results**

### 267 **4.1 Glacial and periglacial evidence**

268 Detailed mapping was conducted in three north-oriented glacial valleys on Mt  
269 Mavrovouni (Figure 4). All valleys extend between the main NW-SE watershed at c. 2100m  
270 a.s.l. and c. 1300-1200m a.s.l where they converge to a main valley. No trace of glaciation  
271 was identified on the southern slopes. The glacial and periglacial features identified in the  
272 three valleys were morphostratigraphically classified upon elevation, relative stratigraphic  
273 position, and degree of degradation (e.g. moraine dismantling) into three distinct units which  
274 are summarised in Table 3. Unit 1 consists of the stratigraphically older glacial features at  
275 relatively low elevations within the valleys characterised by exhumed and relatively eroded  
276 moraines. Unit 2 represents a group of well-preserved moraines with distinct depositional  
277 limits which are limited to higher elevations within the cirques. On the contrary, Unit 3  
278 deposits have been classified as a group of periglacial features (relict rampart ridges) due to  
279 their proximity to the cirque walls, shape and composition (domination of large angular  
280 boulders and lack of fines).

Table 3. Morphostratigraphic classification of glacial and periglacial features on Mt Mavrovouni

Valley	Unit 1	Unit 2	Unit 3
Flega	Lateral Moraine 1740 - 1840m a.s.l.	Scattered glacial boulders Moraine damming a lake 1880-1920m a.s.l.	-
Mnimata	Ice-polished bedrock / Glacial boulders/ Tarns 1900 – 1950m a.s.l.	Terminal and hummocky moraines 2000 – 2040m a.s.l.	Pronival Ramparts 1980 - 2050m a.s.l.
Arkoudolakos (Bear)	Lateral Moraines 1500 – 1660m a.s.l. 1470 – 1510m a.s.l.	Scattered erratic boulders and till with indistinct depositional limits 1800 – 1900m a.s.l.	-



284 *Figure 4. Glacial geomorphological map of the study area on Mt Mavrovouni (Panel A in Figure 1).*

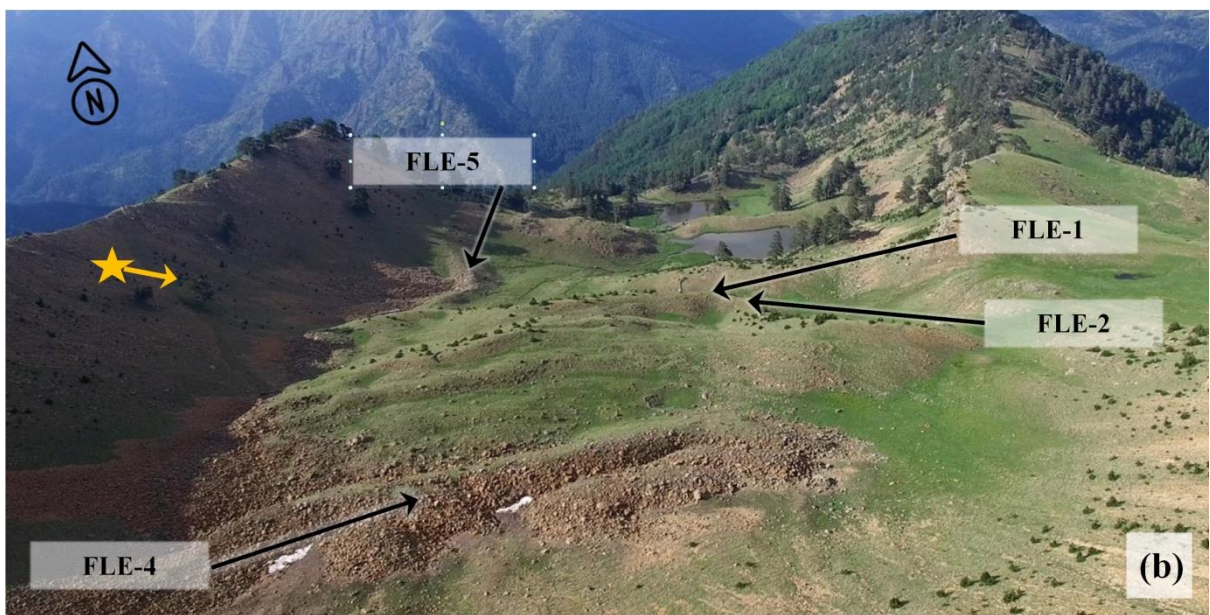
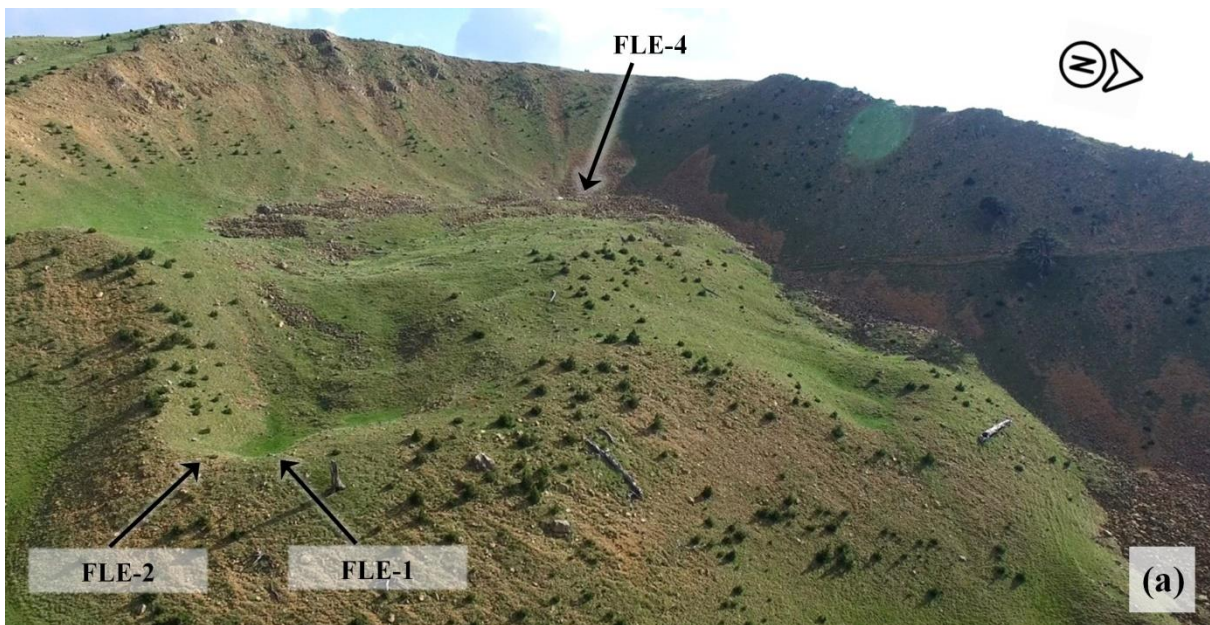
#### 285 **4.1.1 Mnimata (Flega Tarns) Valley**

286 The Mnimata glacial valley extends below the summit of Mt Mavrovouni (Flega Peak at  
287 2157m a.s.l.) with a north orientation (Figure 4). It is typically U-shaped and headed by an  
288 ice-steepened, and well-developed cirque (Figure 5a). A detailed glacial geomorphological  
289 map of the valley is shown in Figure 6 whereas an aerial video captured with an UAV can be  
290 seen [here](#). The upper valley is characterised by a clearly shaped and well-preserved  
291 glacial/periglacial sequence of deposits between 2050 and 2000m a.s.l., which is a focal point  
292 of this research (Figure 5). The lower limit of the sequence is defined by a well-shaped  
293 terminal moraine (Unit 2 in Table 3). Two samples for <sup>36</sup>Cl surface exposure dating were  
294 taken from the moraine crest at 2017m a.s.l. (FLE-1 and FLE-2 in Figure 5a, Figure 6). The  
295 middle part of the sequence is characterised by hummocky moraine with scattered  
296 subrounded glacial boulders that formed upon glacial retreat. These deposits have likely  
297 undergone further permafrost-related creep, though without any typical rock glacier features  
298 as ridges or lobes.

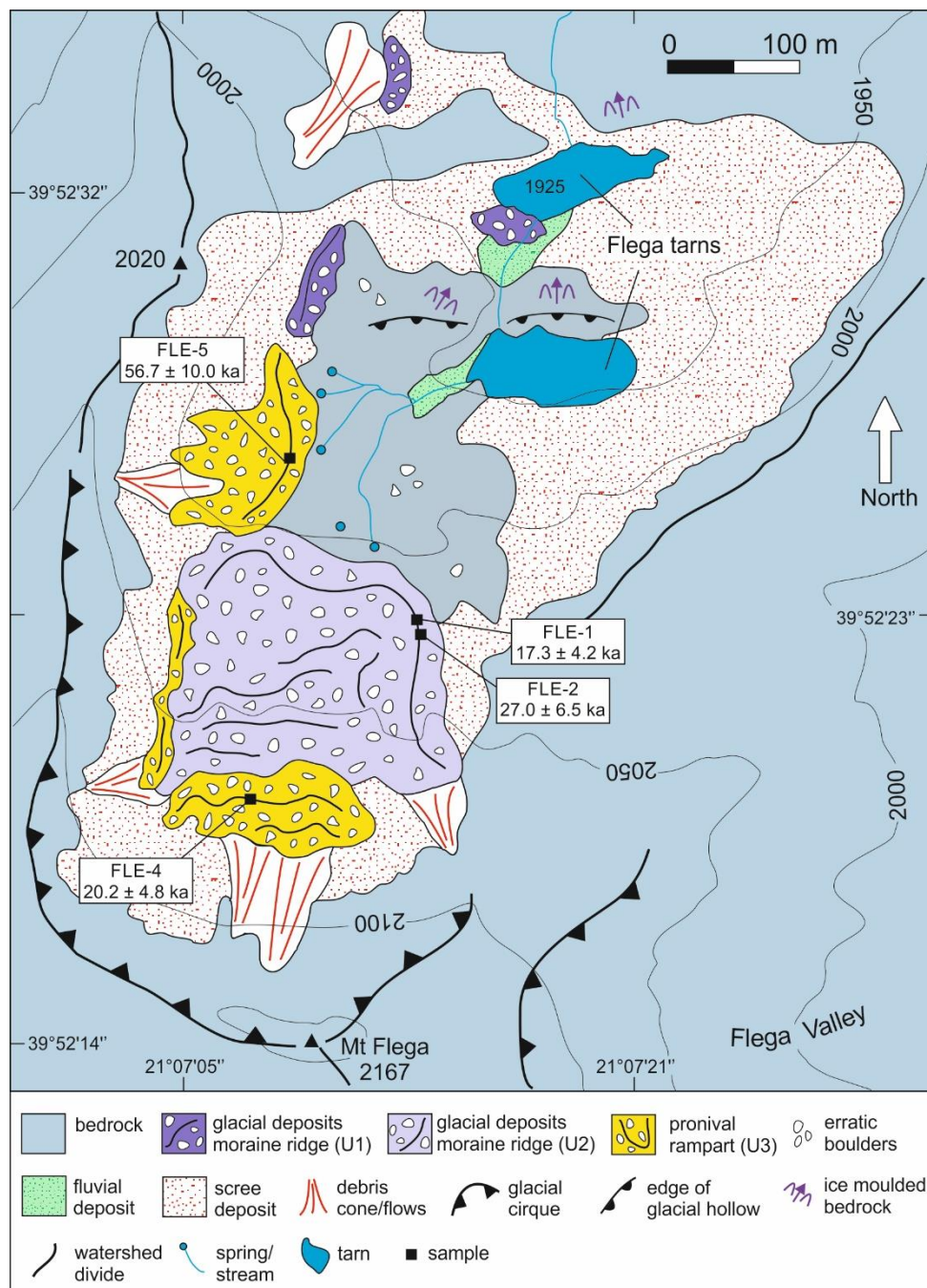
299 Higher in the valley, there is a sharp and continuous boulder ridge with minimum soil  
300 development and mainly composed of large angular boulders that have been deposited at the  
301 centre of the cirque-floor. These boulders are considerably different in shape compared to the  
302 glacially worked and therefore subrounded boulders found lower in the main sedimentary  
303 unit (Unit 2 in Table 3) whereas soil development is minimal compared to the Unit 2  
304 moraines. For these reasons, these deposits are ascribed to a distinct younger stratigraphic  
305 unit (Unit 3 in Table 3, Figure 4 and Figure 6). A sample was taken from a well-stabilised  
306 boulder on the crest of these deposits at 2038m a.s.l. (FLE-4 in Figure 5b, Figure 6). Two  
307 minor crests of similar deposits are nested in the eastern part of the main ridge (Figure 5b).  
308 The main ridge along with the two minor crests of the angular deposits are interpreted as



309 pronival ramparts that formed after the formation of the main sedimentary unit (Unit 2 in  
310 Table 3,) during persisting cold conditions that were too arid for the development of a glacier  
311 but sufficiently humid for the formation of a permanent snow (nevé) field. Local topography  
312 and shading from the cirque-walls must have favoured the formation of this nevé field.  
313 Further up-valley, no boulders or clasts can be found on the grassy scree-covered slope,  
314 supporting the hypothesis of the formation of a pronival rampart during prevailing cold  
315 conditions with increased debris supply due to freeze-thaw weathering of the cirque walls.



318 Figure 5. The upper Mnimata valley and the glacial/periglacial sequence along with sample locations  
 319 for  $^{36}\text{Cl}$  dating: a) (Top): Looking up valley. Notice the Unit 2 terminal moraine delimiting this set of  
 320 deposits in the foreground and the well-developed cirque; b) (Bottom): Looking down valley. Unit 3  
 321 pronival deposits can be seen in the foreground along with the denoted sample location. In the  
 322 background there are the Flega tarns sitting in ice-scoured hollows on the impermeable bedrock.  
 323 Yellow star and arrow denote shooting location and orientation of Figure 7b. (Aerial Photos: George  
 324 Panayiotopoulos, June 2019)





326 *Figure 6. Detailed glacial geomorphological map of the Mnimata Valley. This is panel B in Figure 1.*

327 The middle part of the valley, extending between 2000m a.s.l. and the treeline at 1850m  
328 a.s.l. is characterised by two tarns known as the Flega lakes and is dotted with erratic  
329 boulders (Figure 6, Figure 7a). Along the western flank of the valley, detrital ridges at the  
330 base of the slope were observed. The landform (evident crest slight sinuous or with an  
331 upvalley facing concavity) and the deposit characteristics are pretty similar to those of the  
332 pronival rampart sampled in the upper part of the valley. Therefore, these ridges have been  
333 ascribed to the same morphostratigraphic unit (Unit 3 in Table 3, Figure 4 and Figure 6). A  
334 single sample for  $^{36}\text{Cl}$  dating has been taken from one of the crests of these deposits at 1980  
335 m a.s.l. (FLE-5 in Figure 5b and Figure 7b). Behind the crest there is angular debris that was  
336 most probably deposited after the disappearance of the nevé field according to mass wasting  
337 phenomena.



338



339

340 *Figure 7. a) (Top): Glacially transported and sub-rounded boulders between the two tarns;*

341 *b) (Bottom): The Mnimata Unit 3 pronival rampart deposits at the W flank of the valley. Photo*

342 *location shown in Figure 5b. Sample FLE-5 location is also denoted. Notice the streams nourishing*

343 *the upper Flega tarn that spring below the deposits and onto the ice-scoured exposed impermeable*

344 *bedrock that also hosts the two tarns (Photos August 2016).*

345 The two lakes are glacial in origin (Unit 1 in Table 3) but are not moraine-dammed as

346 initially expected from the analysis of satellite imagery. In fact, they sit on ice scoured

347 hollows on bedrock (Figure 7a). There are at least five springs just above the upper tarn

348 where the aquifer meets the exposed impermeable bedrock (Figure 6, Figure 7b) and these

349 springs nourish the two tarns. Between the tarns there are scattered glacial boulders (Figure

350 7a) whereas a pair of lateral moraines has been preserved on the western flank of the valley

351 between 1900 and 2050m a.s.l. indicating the presence of a significantly larger palaeoglacier

352 compared to the palaeoglacier corresponding to morphostratigraphic Unit 2(Figure 6). Ice-

353 scoured and polished bedrock has been identified just below the lower tarn at 1900m a.s.l.

354 that still bears clear striations in the direction of the paleaoglacier ice-flow on its surface

355 (Figure 3), indicative of low postglacial erosion rates of the rock surface (Gosse and Philips,

356 2001). It should be noted that these features do not represent the lowermost depositional

357 limits of Unit 1 in this valley but rather mark the lower limit of valley accessibility and thus  
358 of the conducted fieldwork. Therefore, the correlation with the lowermost glacial deposits in  
359 the Flega and Arkoudolakos valleys (see Table 3) has been based on their relative  
360 stratigraphic position, palaeoglacier size and erosion extent of the recorded sediments.

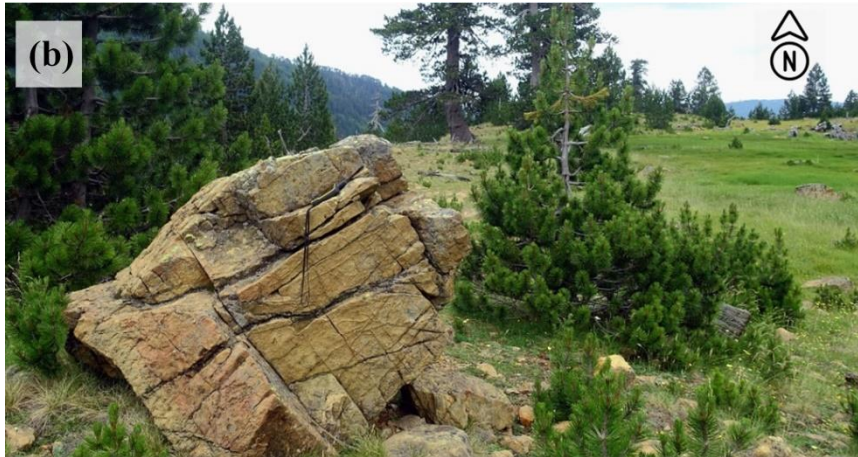
#### 361 **4.1.2 Flega Valley**

362 The Flega glacial valley has a N-NE orientation. Along its main axis runs the  
363 homonymous Flega stream which has caused deep cut-off erosion creating a steep, V-shaped  
364 bank-riverbed profile. The head of the valley is ice steepened, forming a wide and quite well-  
365 developed cirque. The edges of this cirque stand between 1950 and 2050m a.s.l. Glacial  
366 evidence in the upper valley is limited to scattered sub-rounded glacial boulders with quite  
367 indistinct depositional limits. However, a small moraine damming a swamp/seasonal lake  
368 (Figure 8) is present at an altitude of 1880m a.s.l. on the eastern flank of the valley. It has  
369 been correlated with Unit 2 deposits in the Mnimata valley on grounds of altitude and  
370 superficial form and has therefore been ascribed to morphostratigraphic Unit 2 (Table 3,  
371 Figure 4).



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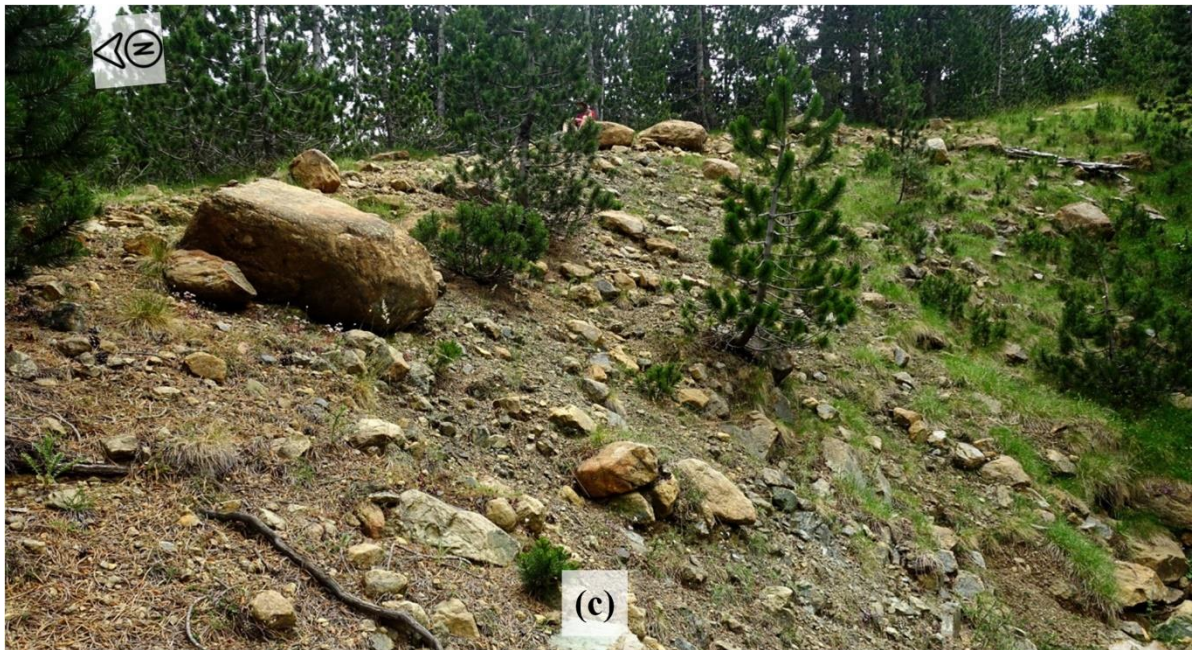
*Figure 8. a) (Top): The Flega Unit 2 moraine which dams a seasonal lake/swamp on the eastern part of the upper Flega valley; b) (Bottom): Glacially transported boulder outstanding on the glacial deposits damming the seasonal lake/swamp at the upper Flega valley. Notice the cracks in the boulder that upon further glacial working or freeze-thaw processes could have comminuted this 1.5m in diameter boulder into smaller clasts. (Photo: July 2018)*



380  
381

*Figure 9. The inner flank of the Flega Unit 1 moraine (Photo: July 2018).*





382 *Figure 10. The Flega Unit 1 moraine; a) (top left): unsorted clast-supported deposits with a fine*  
383 *gravel matrix and sandy to silty material at the lowermost part of the exposure; b) (top right):*  
384 *subrounded clasts of variable size, from a diameter of 25cm to centimetric gravel; c (bottom): larger*  
385 *subrounded and striated glacial boulders along the moraine crest (Photo: July 2018).*



386 In the lower part of the valley (below 1900m a.s.l.), a continuous lateral moraine  
387 extending from 1740 to 1840m a.s.l. in altitude and 600m in length has been classified into  
388 Unit 1 on the grounds of former glacial extent, signs of exhumation and relative stratigraphic  
389 position (Table 3, Figure 4, Figure 9, Figure 10). The moraine converges to the modern river  
390 channel, possibly indicating the position of the former glacier's snout, although the terminal  
391 frontal moraine must have been eroded from run-off water. This moraine has most probably  
392 been formed by a glacier originating from the central-western part of the NE Flega main  
393 cirque and the NE Flega sub-cirque, although an origin from the western section of the cirque  
394 cannot be excluded.

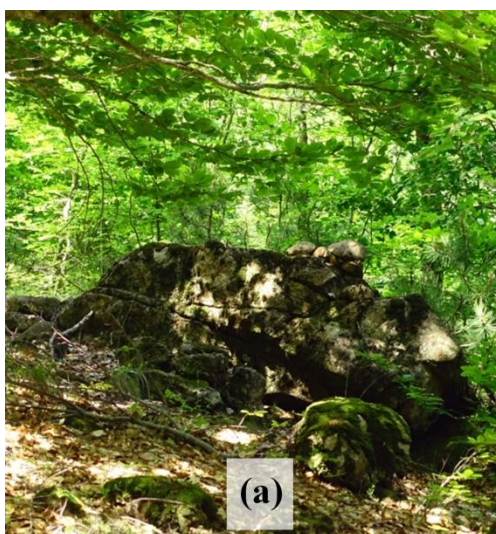
395 These ophiolite-dominated glacial sediments appear to be clast-supported, with  
396 prevalently ophiolitic clasts partly cemented by a scarce sandy to silty matrix (Figure 10a and  
397 b). The clasts are striated and subrounded. Larger boulders do not exceed 1.5m in diameter  
398 and are mostly evident along the moraine crest (Figure 10c). Contrary to moraines of  
399 granite/gneiss or limestone lithologies, no large blocks (>5m in diameter) are present in this  
400 ophiolite-dominated sedimentary unit due to the cracks and joints caused by structural  
401 deformation of the parent bedrock (e.g. Figure 8). These characteristics are consistent with  
402 Middle Pleistocene glacial deposits of ophiolitic lithology from Mt Smolikas (Hughes et al.,  
403 2006c) and Mt Vasilitsa (Hughes, 2004) (Figure 1).

#### 404 **4.1.3 Arkoudolakos valley**

405 The upper part of the Arkoudolakos glacial valley includes two converging valleys  
406 (Figure 4). The western branch has a NE orientation and is quite steep and inaccessible, thus  
407 it was excluded from field work. The N-oriented eastern branch is a U-shaped and wide  
408 valley but its head is smooth and does not bear typical glacial geomorphological features like  
409 a cirque or ice-steepened headwalls. However, scattered glacial boulders and remnants of  
410 sub-glacial till are evident throughout this valley branch (Figure 4). Similarly to the Flega

411 valley, the depositional limits are indistinct and, although these deposits could be correlated  
412 on terms of altitude with morphostratigraphic Unit 2 deposits from the two adjacent valleys,  
413 they could also very well be synchronous to the older and more extensive deposits lower in  
414 the valley that belong to morphostratigraphic Unit 1 (Figure 4).

415 The lower Arkoudolakos valley (part below the converging point of the two upper  
416 branches) has a N orientation (Figure 4), is densely forested and along its central axis has  
417 undergone V-shaped erosion by run-off water, which is typical for ophiolites (Hughes et al.,  
418 2006d). Two moraines have been identified in the lower part of the main valley forming the  
419 Bear Unit 1 moraines (Table 3). The first is a clear and well-shaped lateral moraine that  
420 extends right from the point where the two upper branches of the valley converge at 1660 m.  
421 a.s.l. and down to 1500m a.s.l. (Figure 4) where it has been washed away by the river that  
422 runs along the valley (Figure 11). Its total length is about 800 m. The second moraine is  
423 poorly preserved, extends from 1510 to 1470m a.s.l., and is 300m long. Both moraines are  
424 covered by dense beech forests and thus are unrecognisable from satellite imagery. No  
425 exposure of the moraines were discovered. The Bear Unit 1 moraines are correlated with the  
426 Flega Unit 1 moraine in the Flega valley on terms of altitude, stratigraphic position and signs  
427 of erosion (Figure 4)



428



429

430 *Figure 11. The Bear Unit 1 moraine; a) (up left): large glacial boulder (d~2 m) along the densely*  
 431 *forested moraine b) (up right): subrounded and striated boulders and clasts along the well-shaped*  
 432 *moraine crest and; c (below): another glacially transported large boulder next to the path that runs*  
 433 *along the moraine ridge (backpack length ~0.6m for scale).*

#### 434 4.2 Geochemistry

435 The geochemistry of the samples is presented in Table 4. All analyses were conducted by  
 436 the Acme labs in Canada (Bureau Veritas Mineral Laboratories). Bulk rock compositions  
 437 refer to the sample prior to any etching or cleanup, while the target rock sample represents  
 438 the final, clean mineral fraction that was dissolved for AMS analysis. Bulk rock compositions  
 439 are used to calculate neutron-transport parameters, which are particularly important for  
 440 samples with significant production from neutron capture (see Marrero et al., 2016b).

441 *Table 4. Geochemistry of the serpentized peridotite samples for <sup>36</sup>Cl dating. For major and minor*  
 442 *elements, results are given for both bulk rock and target (processed) rock samples.*  
 443 *(\* ICP-ES, \*\*ICP-MS, \*\*\*LiBO<sub>2</sub>/Li<sub>2</sub>B<sub>4</sub>O<sub>7</sub> fusion & ICP-ES)*

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Major and minor element mass fraction (wt %)

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Sample ID	SiO <sub>2</sub>	Al <sub>2</sub> O <sub>3</sub>	Fe <sub>2</sub> O <sub>3</sub>	MgO	CaO	Na <sub>2</sub> O	K <sub>2</sub> O	TiO <sub>2</sub>	MnO	Water (LOI)	CO <sub>2</sub>
Bulk rock samples*											
FLE-1	43.44	0.43	8.21	41.22	0.36	0.02	0.01	<0.01	0.11	4.6	0.6
FLE-2	44.17	0.22	8.48	42.33	0.24	0.01	<0.01	<0.01	0.12	2.9	0.6
FLE-4	41.48	0.21	8.35	40.38	0.21	0.01	<0.01	<0.01	0.11	7.5	0.6
FLE-5	45.03	1.61	8.35	39.45	1.42	0.03	<0.01	0.02	0.12	2.5	0.6
Target rock samples**											
FLE-1	45.98	0.43	8.63	39.36	0.39	0.01	<0.01	<0.01	0.11	-	-
FLE-2	46.3	0.22	8.88	41.26	0.24	<0.01	<0.01	<0.01	0.12	-	-
FLE-4	44.16	0.24	8.58	39.16	0.2	<0.01	<0.01	<0.01	0.1	-	-
FLE-5	46.21	1.8	8.64	38.62	1.61	0.03	<0.01	0.02	0.12	-	-
Trace element concentrations (ppm)											
Sample ID	B	Sm	Gd	U	Th	Li	V	Ba	Ni	Sc	Co
Bulk rock***											
FLE-1	<3	<0.05	<0.05	<0.1	<0.2	<0.001	22	8	2388	6	123.6
FLE-2	<3	<0.05	<0.05	<0.1	<0.2	<0.001	16	10	2473	5	126.5
FLE-4	<3	<0.05	<0.05	<0.1	<0.2	<0.001	23	4	2470	4	123.6
FLE-5	<3	<0.05	0.06	<0.1	<0.2	<0.001	51	6	2214	11	113.3

444 Bulk rock compositions primarily reflect the proportions of minerals present. Based on  
445 these compositions, these rock samples are peridotites/harzburgites with predominant  
446 serpentine alteration (Dilek et al., 1997a; Kelemen et al., 2004). A rock density of 2.7 g cm<sup>-3</sup>  
447 was considered for all samples (Table 1), in accordance to the respective values for  
448 ultramafic peridotites with medium to high serpentinization degree (>70%) from the Pindus  
449 ophiolitic complex (Bonnemains et al., 2016). XRD analyses of highly serpentinized  
450 ultramafic rocks from oceanic crust in the Mid-Atlantic Ridge showed that Loss On Ignition  
451 (LOI) can be attributed entirely to water content (Kelemen et al., 2004) whereas an average  
452 value for CO<sub>2</sub> concentrations is 0.6 wt% (Kelemen et al., 2011). It is noted that the slight

453 underestimation of the total volatile content is caused by the conversion of FeO to Fe<sub>2</sub>O<sub>3</sub>  
454 during heating of the sample powders for LOI estimation (Kelemen et al., 2004). This  
455 approach was adopted in this paper, attributing the measured LOI entirely to water content  
456 and assuming a 0.6 wt% of CO<sub>2</sub> concentration for all samples (Table 4).

#### 457 4.3 <sup>36</sup>Cl exposure ages

458 The two samples FLE-1 and FLE-2 were taken from the largest (up to 1m diameter) and  
459 most stable boulders which were partly embedded along the ridge of a well-consolidated  
460 terminal moraine (Figure 12a) reducing the possibility of dating inaccuracies due to post  
461 depositional processes such as moraine dismantling and boulders toppling . FLE-1 and FLE-2  
462 yielded <sup>36</sup>Cl exposure ages of 17.3 ± 4.2 ka and 27.0 ± 6.5 ka, respectively.





465 *Figure 12. a) (Top): Sampled boulder on the well-consolidated terminal moraine (samples*  
 466 *FLE-1 & FLE -2); b) (Bottom): Sampled boulder on the crest of the less-consolidated*  
 467 *pronival deposits (sample FLE-4). Exact samples locations are given in Figures 5 and 6.*  
 468 *(Photos: October 2016)*

469 Further upvalley, pronival deposits are less consolidated than the morainic deposits and  
 470 are typically characterised by larger (1-5m diameter) and more angular boulders and the  
 471 absence of a matrix consisting of gravels and fines (Hedding and Sumner, 2013). Two stable  
 472 boulders along the most consolidated part of the pronival deposits were chosen for sampling  
 473 (FLE-4 and FLE-5, Figure 12b). The  $^{36}\text{Cl}$  exposure ages for these two samples were  
 474 calculated at  $20.2 \pm 4.8$  ka and  $56.7 \pm 10.0$  ka respectively.

475 The apparent exposure ages of the four samples as well as the respective elemental  
 476 contribution to total  $^{36}\text{Cl}$  production as calculated with the CRONUScalc  $^{36}\text{Cl}$  Exposure Age  
 477 Calculator v2.2 are summarised in Table 5. All ages are within the Late Pleistocene as  
 478 initially expected but most importantly samples FLE-2, FLE-4 and FLE-5 fall in the LGM  
 479 interval within the respective uncertainties.

480 *Table 5. Exposure ages, share of total  $^{36}\text{Cl}$  production per element as calculated with CRONUScalc*  
 481  *$^{36}\text{Cl}$  Exposure Age Calculator v2.2 for the considered samples. Uncertainties include analytical and*  
 482 *production rate uncertainties. Reported percentages of  $^{36}\text{Cl}$  production correspond to production from*  
 483 *spallation and slow muons for Ca and K, low energy neutrons for Cl and spallation only for Fe and*  
 484 *Ti.*

<i>Results obtained with CRONUScalc <math>^{36}\text{Cl}</math> Exposure Age Calculator v2.2</i>						
Sample ID	Age (ka)	Percentage of total $^{36}\text{Cl}$ production (%)				
		Cl	Ca	K	Fe	Ti
FLE-1	$17.3 \pm 4.2$	88.23	6.11	0	5.65	0
FLE-2	$27.0 \pm 6.5$	91.02	3.52	0	5.45	0
FLE-4	$20.2 \pm 4.8$	96.97	1.08	0	1.95	0

FLE-5	56.7 ± 10.0	56.94	35.22	0	7.80	0.04
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486 **4.3.1 Implications of iron-rich ophiolitic lithologies on cosmogenic <sup>36</sup>Cl exposure**  
 487 **dating**

488 The glacial chronology of Mt Mavrovouni relies on the application of cosmogenic <sup>36</sup>Cl  
 489 surface exposure dating on highly serpentinized ophiolites. In this study, due to low Ca and K  
 490 concentrations, the cosmogenic <sup>36</sup>Cl has been mainly produced from Cl with a non-negligible  
 491 contribution from Fe (Table 5), making these production rates more important than in  
 492 samples of other lithologies. The relatively high uncertainties of the calculated ages are a  
 493 result of the domination of <sup>36</sup>Cl production from the capture of thermal neutrons by <sup>35</sup>Cl in  
 494 the rock and the higher uncertainty associated with the production from this pathway  
 495 (Marrero et al, 2016a; Asimov and Ivy-Ochs, 2009).

496 The Fe pathway accounts for ~2-8% of the total <sup>36</sup>Cl production (Table 5), which is much  
 497 higher than the 1-2% combined Ti and Fe contributions in most common rock types (Marrero  
 498 et al., 2016a), so it has not been included as a calibrated parameter in recent calibrations (e.g.  
 499 Schimmelpfennig et al., 2009; 2011; Marrero et al., 2016a ). In our case, the rock samples are  
 500 characterised by high concentrations of Fe<sub>2</sub>O<sub>3</sub> (8-8.5%) (Table 4). As the current version of  
 501 the CRONUScalc online calculator (v2.1;  
 502 <http://cronus.cosmogenicnuclides.rocks/2.1/html/cl/>; Marrero et al., 2006b) has not been  
 503 specifically calibrated for production rates from Fe spallation, significant production from Fe  
 504 was investigated further here. The exposure ages of the samples were recalculated using a  
 505 modified version of the original CRONUScalc code (v2.2, available on  
 506 <https://bitbucket.org/cronusearth/cronus-calc/>; accessed 15 Sept 2021). In this version, new  
 507 Fe production rates of 1.29 ± 0.11 atoms <sup>36</sup>Cl (g Fe)<sup>-1</sup> yr<sup>-1</sup> as reported from Moore and  
 508 Granger (2019) are incorporated into CRONUScalc and used in conjunction with the LSDn  
 509 scaling model as described in Anjar et al. (2021). This required recalibrating the main <sup>36</sup>Cl

510 production rates, resulting in <1% change in the production rates for the main pathways  
511 (Anjar et al., 2021) when compared to the previous production rates (Marrero et al., 2016a;  
512 2021). The results obtained show negligible differences (<1%) in the resulting ages compared  
513 with the results obtained with the latest online version (v2.1) of the calculator. Moore and  
514 Granger (2019) underline that additional data are needed to confirm the scaling behaviour  
515 described in their work, so high-Fe samples like these could be used to contribute towards  
516 this effort.

## 517 **5 Discussion**

### 518 **5.1 Glacial chronology and interpretation**

519 Sampling was limited to the stratigraphically youngest set of glacial deposits in the  
520 Mnimata valley (Figure 6). The sampling strategy was dictated by the aim to establish a  
521 preliminary chronology of this glacial/periglacial sequence and confirm that its formation  
522 falls within the LGM as suggested by morphostratigraphic correlations of moraines within the  
523 Pindus mountains (Mt Tymphi, Mt Smolikas and Mt Mavrovouni; Figure 1). However, the  
524 small number of ages and the high level of uncertainty in the calculated ages means that  
525 additional data are needed in order to establish a tighter and reliable chronology of cold  
526 events on Mt Mavrovouni. Next, a preliminary chronology according to the obtained  $^{36}\text{Cl}$   
527 surface exposure ages is presented.

528 The  $^{36}\text{Cl}$  exposure ages for the main terminal moraine samples FLE-1 and FLE-2 (Unit 2  
529 in Table 3, Figure 4 and Figure 6) were  $17.3 \pm 4.2$  ka and  $27.0 \pm 6.5$  ka, respectively (Table  
530 5). The older age is considered to be the most representative for moraine deposition whilst the  
531 younger one may represent a period of boulder toppling or exhumation (Ivy-Ochs et al.,  
532 2006a, 2006b, 2007; Ivy-Ochs and Kober, 2008; Ivy-Ochs and Briner, 2014). This is because  
533 sample FLE-1 (which yielded the younger age) was taken from the proximal (inner) side of  
534 the moraine which is more susceptible to degradation and mass wasting processes. FLE-2

535 (which yielded the older age) was taken from one of the highest points on the moraine ridge  
536 making it less likely to have undergone significant toppling or exhumation (Ivy-Ochs et al.,  
537 2006a, 2006b, 2007). Assuming the mean of the two ages as landform age was not considered  
538 as it is susceptible to significant inaccuracies (Ivy-Ochs et al., 2007), especially if incomplete  
539 exposure and moraine dismantling are suspected (Applegate et al., 2010; D’Arcy et al.,  
540 2019). Thus, in this preliminary chronology this moraine is considered to have been formed  
541 by an advancing cirque glacier when it reached its maximum extent close in time to the  
542 global LGM interval (27.5-23.3 ka; Hughes and Gibbard, 2015) as the age of sample FLE-2  
543 falls within this interval but additional data is in any case needed to strengthen this  
544 chronology. This age also provides the maximum age of formation of Unit 2 glacial deposits  
545 on Mt Mavrovouni which falls close to the LGM interval within the calculated age  
546 uncertainties.

547 As the geomorphological evidence (terminal moraine and cirque walls) define sufficiently  
548 the limits of this palaeoglacier it was possible to use a geographical information systems  
549 approach (GIS) to calculate its palaeo-ELA at 2090m above modern sea level. In particular,  
550 the surface, thickness and ELA of the former glaciers were reconstructed using the semi-  
551 automated GIS tools based on the numerical technique of Benn and Hulton (2010) and  
552 developed by Pellitero et al. (2015; 2016). The ELA was calculated with the adaptation of the  
553 classic area– altitude balance ratio (AABR) method (Osmaston, 2005) using a balance ratio  
554 (BR) of 1.6, which is the average obtained on present-day glaciers in other Mediterranean  
555 mountains (Rea, 2009).

556 The stratigraphically youngest deposits within the Mnimata sequence (Unit 3 in Table 3,  
557 Figure 4 and Figure 6) are interpreted as relict pronival ramparts. More importantly, their  
558 preliminary age of deposition discussed here marks a cold period with persistent snow/ice  
559 cover or permafrost occurrence whilst at the same time constitutes the lower limit for the

560 timing of deposition of the Mnimata sequence. The uppermost part of this set of deposits  
561 (FLE-4) and that at the western flank of the upper valley (FLE-5) were expected to have  
562 formed synchronously regardless of their different altitude as they belong to two  
563 morphologically and compositionally similar ridges expanding along the W-E and S-N axes  
564 respectively and both have been deposited directly below the slopes above them. However,  
565 the samples yielded ages of  $20.2 \pm 4.8$  ka and  $56.7 \pm 10.0$  ka respectively. The most  
566 representative age for the formation of these features is considered to be the younger one as  
567 the pronival ramparts in question are stratigraphically younger than the terminal moraine  
568 ridge that was dated at  $27.0 \pm 6.5$  ka. Moreover, the FLE-4 sample was taken from a large  
569 and well stabilised boulder on the highest point of the pronival rampart within a section of  
570 angular deposits lacking finer material and therefore minimising the probability of  
571 exhumation or other mass wasting processes (Allard et al., 2020). On the other hand, the  
572 FLE-5 age could have been affected by inheritance of  $^{36}\text{Cl}$  atoms from previous exposure.  
573 However, marked inheritance in moraine boulders (and similarly in pronival ramparts) is rare  
574 (Heyman et al., 2011) and incomplete exposure is more likely to be an issue (Ivy-Ochs and  
575 Briner, 2014). Therefore, a more probable interpretation is that the morphology of this  
576 sedimentary unit is composite. It is likely that an older lateral moraine associated with the  
577 stratigraphically older Unit 1 pre-LGM deposits was later partly overlapped by a nested  
578 pronival rampart. In this case, the moraine itself became, once the glacier had disappeared, a  
579 starting ridge at the base of the slope which then evolved into a pronival rampart. The age of  
580 this sample could then represent a period of boulder exhumation. Additional dating data are  
581 also needed in this case to support this preliminary chronology. The deposition age of Unit 3  
582 sediments are is thus here considered at  $20.2 \pm 4.8$  ka, probably within the LGM interval but  
583 with a considerable level of uncertainty as the respective age range indicates.



584 The most extensive glaciation phase is represented by a single morphostratigraphic unit  
585 (Unit 1 in Table 3, Figure 4 and Figure 6). This unit may represent more than one phase of  
586 glaciation, possibly widely separated in time, but it was not possible to distinguish multiple  
587 moraine units that correspond to different glacial stages, possibly widely separated in time,  
588 but it was not possible to distinguish multiple moraine units that correspond to different  
589 glacial stages. Nevertheless it is clearly distinguished from Units 2 and 3 as these  
590 stratigraphically older moraines are at lower altitudes and show considerably more signs of  
591 erosion and higher level of soil development. These moraines remain undated but, on the  
592 basis of morphostratigraphic correlations (relative stratigraphic position, elevation, erosion  
593 and soil development) with the moraines in the nearby Mt Tymphi (Woodward et al., 2004;  
594 Hughes et al., 2006a) as well as in Mt Chelmos in southern Greece (Pope et al., 2017; Figure  
595 1) and in the mountains of Montenegro (Hughes et al. 2010, 2011), they are likely Middle  
596 Pleistocene in age. Particularly, this glacial phase may correspond to the Skamnellian  
597 (MIS12) and/or the Vlasian Stage (MIS6) of northern Greece.

## 598 **5.2 The Glacial history of the Pindus Mountains and Greece in a regional context**

### 599 **5.2.1 Pre-LGM Pleistocene glaciations**

600 The most extensive glacial phases in the mountains of Greece most probably took place  
601 during Middle Pleistocene and particularly during MIS 12 (Skamnellian Stage) and MIS 6  
602 (Vlasian Stage). This was established by the studies on Mt Tymphi (Woodward et al., 2004;  
603 Hughes, 2004; Hughes et al., 2006a) and in the mountains of Montenegro (Hughes et al.,  
604 2010, 2011) based on maximum U-series ages from secondary calcites present in limestone-  
605 derived moraines/till. It should be noted that the ages of the Skamnellian Stage moraines are  
606 beyond the upper limit of U-series dating which is 350 ka (Woodward et al., 2004; Hughes et al.,  
607 2006a). It is important to recognise that as U/Th ages denote the age of secondary calcites  
608 formation sometime after deposition and in this case during the more favourable warm

609 interglacial periods, the oldest ages for a particular deposit can be used to infer glacial activity  
610 during a previous cold stage (Woodward et al., 2004). For this reason, the ages from the older  
611 moraines (>350 ka) indicate that calcites formation was associated with MIS 11 (or older)  
612 although the precise ages are unknown (Woodward et al., 2004; Hughes et al., 2006a). Taking  
613 also into account that the next glacial period preceding 350ka (MIS 11) is MIS 12 which is  
614 considered to be the most extreme climatic glacial interval of the last ~500 ka (Tzedakis et al.,  
615 2003) the deposition of these moraines was ascribed to this interval (Hughes et al., 2006a).

616 This chronology could potentially be independently tested with the application of  $^{36}\text{Cl}$   
617 surface exposure dating to the ice-scoured bedrock recorded below the twin tarns in the  
618 Mnimata valley in order to constrain the timing of the most extensive glacial phase on Mt  
619 Mavrovouni (Unit 1 in Table 3; Figure 3). However, it is likely that exposure dating on such  
620 old surfaces may yield just minimum ages due to surface erosion, although it should be a  
621 target for future research.

### 622 **5.2.2 Glaciations during the global LGM**

623 The LGM  $^{36}\text{Cl}$  dates presented in this paper from ophiolite-derived moraines on Mt  
624 Mavrovouni are consistent with the  $^{36}\text{Cl}$  dates from limestone moraines on the nearby Mt  
625 Tymphi (Allard et al., 2020), suggesting a Late Pleistocene local glacial maximum close in  
626 timing to the global LGM interval (27.5-23.3 ka) within the errors of the calculated ages. In  
627 particular, the largest Late-Pleistocene glaciers on Mt Tymphi reached their terminal  
628 positions no later than  $25.7 \pm 2.6 - 29.0 \pm 3.0$  ka and had an ELA of 2016m a.s.l. (Allard et al.,  
629 2020). Glaciers had retreated to the high cirques by  $24.5 \pm 2.4$  ka (Allard et al., 2020) during  
630 Heinrich Stadial 2 (HS2: 26.5 -24.3 ka; Sanchez Goñi and Harrison, 2010). This timing of a  
631 near-LGM late Pleistocene glacier maximum is confirmed by the  $^{36}\text{Cl}$  geochronology of  
632 ophiolitic glacial boulders from a terminal moraine on Mt Mavrovouni presented here,  
633 indicating stabilisation of the most extensive Late Pleistocene glaciers at  $27.0 \pm 6.5$  ka. On

634 nearby Mt Smolikas (Figure 1), the well-formed moraines at 2000-2200m a.s.l., which have  
635 been ascribed to the last cold stage (MIS5d-2) but remain undated (Hughes et al., 2006d),  
636 may also be attributed to a glacial advance phase prior to, or within, the LGM.

637 The timing constraint of a local Late Pleistocene glacial maximum in the Pindus  
638 mountains in northwest Greece during the LGM interval is in good agreement with well-  
639 preserved outwash sediments dating to  $28.2 \pm 7.0 - 24.3 \pm 2.6$  ka (U-series; ESR, TL ages;  
640 Lewin et al., 1991; Hamlin et al., 2000; Macklin and Woodward, 2009 and references  
641 therein) in the Voidomatis River catchment which drains the southern part of Mt Tymphi  
642 including extensively glaciated slopes. This is also consistent with the Ioannina basin pollen  
643 record about 30km southwest from Mt Mavrovouni (Tzedakis et al., 2002; Figure 1),  
644 indicating cool and wet conditions, most favourable for glacier growth, at 30-25 ka (see  
645 discussion in Allard et al., 2020). The late half of the LGM interval (25.5 - 23.3 ka) is less  
646 favourable for a local glacial maximum despite coinciding with climatic deterioration during  
647 HS2 and a marked regional temperature depression between 25.6 and 23.2 ka (Macklin et al.,  
648 1997; Galanidou et al., 2000; Hughes, 2004). This is because cold and drier conditions  
649 reduced regional moisture supply inhibiting glacial growth (see discussion in Allard et al.,  
650 2020).

651 In the Peloponnese in southern Greece, Pope et al. (2017) showed that Late Pleistocene  
652 glaciers on Mt Chelmos (Figure 1) reached their maximum extent at  $36.5 \pm 0.9 - 28.6 \pm 0.6$   
653 ka with a mean ELA of 2046 m a.s.l. (Pope et al., 2017 -recalculated ages by Allard et al.,  
654 2020) suggesting that glaciers from MIS 3 to 2 in the cirques of Mount Chelmos are likely to  
655 have oscillated with varying climatic conditions, especially in response to millennial-scale  
656 shifts between cold/dry and cool/wet conditions in Greece (Leontaritis et al., 2020; see also  
657 discussion in Hughes et al., 2006c). The results here ( $27.0 \pm 6.5$  ka) cannot be used to rule out  
658 a pre-LGM advance within the Late Pleistocene, taken into consideration that this age is more

659 likely to represent a minimum age of exposure after the glacier had retreated from the  
660 moraine (and especially given the assumption of negligible erosion in the age calculation).  
661 However, given the limited sample size and although the presented chronology is well  
662 supported by local and regional morphostratigraphic correlations, more data is needed to  
663 build a more robust age model.

664 As regards the calculated ELAs of LGM palaeoglaciers across Greece a crucial issue when  
665 attempting comparisons is the consideration of the tectonic uplift history of the different massifs  
666 (Giraudi and Giaccio, 2015). The North Pindus mountains are characterised by estimated  
667 tectonic uplift rates of 0.4–0.8 mm/yr corresponding to a total uplift of 10-20m for the last 25ka  
668 although rates might be higher (King and Bailey, 1985). These roughly estimated rates regard a  
669 wide area (>9000 km<sup>2</sup>) with many local faults in the numerous massifs and include significant  
670 uncertainties. (Leontaritis, 2021). It is therefore likely that uplift rates may vary for each  
671 mountain area and especially between the limestone-dominated Mt Tymphi and the ophiolitic Mt  
672 Mavrovouni. Thus, any conclusions derived from ELA comparisons between these massifs  
673 may be problematic. Tectonic movements are even more active in the northern Peloponnesus  
674 area, where uplift rates have been estimated at 1.1-1.3 mm/yr over the last 350-500ka  
675 corresponding to a total uplift of 37-40m for the last 25ka (Armijo et al., 1996; De Martini et al.,  
676 2004; McNeill and Collier, 2004).

677 The first estimates of ELAs of LGM palaeoglaciers on Mt Mavrovouni (2090m a.s.l.), Mt  
678 Tymphi (~2000m a.s.l.) and Mt Chelmos (~2010m a.s.l.) indicate that glaciers during the  
679 LGM may have formed under similar climatic conditions across Greece. However, given the  
680 uncertainties in the uplift history of these mountains this conclusion should be considered with  
681 caution.

682 A pre-LGM local Late Pleistocene glacier maximum is evidenced in other Mediterranean  
683 mountains (Leontaritis et al., 2020). In particular, similar pre-LGM glacier advance phases



684 have been recorded in the Italian Apennines (33-27 ka; Giraudi, 2012; Giraudi and Giaccio,  
685 2015 and references therein), the Pyrenees (García-Ruiz et al., 2010), the Segundera and  
686 Cabrera mountain ranges in northern Spain (33 ka; Rodriguez-Rodriguez et al., 2011), the  
687 Sierra Nevada in southern Spain (30-35 ka; Gómez-Ortiz et al., 2012; Palacios et al., 2016)  
688 and the western and central Taurus Range in Turkey [ $35 \pm 2.5$  -  $28.1 \pm 2.6$  ka on Mt Akdağ  
689 (Sarıkaya et al., 2014);  $46.0 \pm 7$  -  $29.8 \pm 2.3$  ka on Mount Bolkar (Çiner and Sarıkaya, 2017);  
690  $29.7 \pm 2.9$  ka on the Dedegöl Mountains (Köse et al., 2019). Glacial advance during  
691 MIS3/early MIS2 in these areas has been associated with the prevalence of optimal cold and  
692 wet conditions (Oliva et al., 2019).

### 693 **5.2.3 After the global LGM**

694 At a global level, the LGM is succeeded by an interval associated with changes on the  
695 northern summer insolation induced by orbital forcing that resulted in the onset of  
696 Termination I at high latitudes and mountain regions at 19–20 ka, as well as an abrupt rise in  
697 sea level (Clark et al., 2009; Shakun et al., 2015). On Mt Mavrovouni in Greece, the  
698 presented evidence from periglacial features (Unit 3 in Table 3; FLE-4 in Table 5 and section  
699 5.1) suggests that after the LGM, cold conditions persisted until  $20.2 \pm 4.8$  ka but were  
700 unable to sustain dynamic glaciers. The deposition of pronival ramparts has been associated  
701 with the presence of perennial snow/nevé fields within the topographically shaded glacial  
702 cirque slopes and the increased debris supply due to freeze-thaw weathering of the cirque  
703 walls. Furthermore, on Mt Tymphi, initial results suggest that small glaciers ( $<0.6$  km<sup>2</sup>)  
704 persisted only in the topographically shaded northeast cirques up until  $18.0 \pm 1.9$  ka, (Allard  
705 et al., 2020). The presence of rock glaciers below these small cirque glaciers, which are  
706 stratigraphically younger than the cirque moraines formed at  $25.7 \pm 2.6$  ka (Allard et al.,  
707 2020), has been interpreted as the result of arid and cold climate conditions in the interval 25-  
708 20 ka (Hughes et al., 2003). The evidence from Mt Mavrovouni and Mt Tymphi thus suggests

709 unfavourable (cold and arid) conditions for glacier development during this transitional phase  
710 from the LGM to the Late-glacial but cold/humid enough to preserve perennial snow fields  
711 (i.e. nevé fields), permafrost and rock glaciers as well as small glaciers in topographically  
712 favourable positions. This is consistent with other proxies derived from Lake Ioannina –  
713 approximately 30 km south of Mt Tymphi and Mt Mavrovouni (Figure 1) – which indicate  
714 locally cold and drier climate persisting during this period that would have inhibited glacier  
715 development (see discussion in Allard et al., 2020), such as low lake levels between 22 and  
716 20 ka (Frogley, 1998) and a drop of temperatures by 7-10°C with annual rainfall around 600  
717 mm as it has been estimated by the analysis of the lacustrine pollen record (Tzedakis et al.,  
718 2002). Finally, a decline in limestone-derived fine sediment input from  $24.3 \pm 2.6$  to  $19.6 \pm$   
719  $3.0$  ka (U-series ages) in the Voidomatis River record implies shrinking glaciers in its  
720 headwaters in the south-facing cirques of Mt Tymphi compared to the earlier part of the  
721 LGM (Macklin et al., 1997). The younger age limit is consistent with unfavourable  
722 conditions for glacial growth and the deposition of pronival ramparts on Mt Mavrovouni and  
723 overlaps within uncertainty with a possible moraine deposition in the northeast cirques of Mt  
724 Tymphi discussed above, thus giving some support to this initial evidence for small  
725 glaciers/neves in the 20-18 ka interval (Allard et al., 2020). Nonetheless, more data are  
726 needed from this period in the Pindus Mountains in order to build a more robust age model.

#### 727 **5.2.4 The Pindus Mountains: a summary**

728 The recently produced geochronological data for the Last Glacial Cycle (MIS 5d-2) in the  
729 mountains of Greece have proved crucial for addressing the most important temporal gap in  
730 the geochronology of the glacial history of Greece (Leontaritis et al., 2020). The Late  
731 Pleistocene  $^{36}\text{Cl}$  surface exposure ages have been carefully examined in the current glacial  
732 framework of the Pindus Mountains following recommended approach of Ivy-Ochs and  
733 Kober (2008). This includes geomorphological field studies, local and regional

734 morphostratigraphic correlations (Palmentola et al., 1990; Boenzi et al., 1992; Woodward et  
735 al., 2004; Hughes, 2004; Hughes et al., 2006a; 2006d; Allard et al., 2020) as well as  
736 independent age constraints (U-series - Hughes et al., 2006a;  $^{36}\text{Cl}$  dating of limestone-  
737 dominated moraines - Allard et al., 2020;  $^{36}\text{Cl}$  dating of ophiolite-dominated moraines - this  
738 paper). However, although the Late Pleistocene glacial record of the Pindus mountains in  
739 Greece is today one of the best-established and most accurately dated in the Balkans, the  
740 existence and dynamics of glaciers in the Pindus mountains during the cold events of the  
741 Late-glacial (17.5-11.7 ka; Rasmussen et al., 2006, 2014) remains poorly elucidated. Late-  
742 glacial features in the Pindus mountains are absent both on Mt Mavrovouni and on Mt  
743 Tymphi (Allard et al., 2020) and their presence seems to be limited to Mt Smolikas (Hughes  
744 et al. 2006d; Leontaritis et al., 2020), where four discrete phases of Pleistocene glacial  
745 activity are recognised (MIS12; MIS6; LGM and post-LGM; Hughes et al., 2006d). The Mt  
746 Smolikas post-LGM glacial phase might correspond to the Younger Dryas (Hughes, 2004;  
747 Hughes et al., 2006d). The glacial sequence on Mt Smolikas is the most complete recorded  
748 glacial sequence in the Pindus mountains (Leontaritis et al., 2020) but its potential has only  
749 been partially explored as its ophiolitic lithology has prevented (until now) the dating of its  
750 moraines with absolute dating methods (Hughes et al., 2006a; Woodward and Hughes, 2011).  
751 The applicability of the cosmogenic  $^{36}\text{Cl}$  surface exposure dating method to ophiolites  
752 demonstrated here opens up the possibility of a future dating program on Mt Smolikas.

## 753 **6 Conclusions**

754 In addition to providing important new insights to the glacial geomorphology of the  
755 Pindus Mountains this paper presents new surface exposure age results from a pilot study  
756 using  $^{36}\text{Cl}$  to date ophiolite glacial boulders. Despite the small number of samples and  
757 relatively large uncertainties, these ages confirm that the last glaciation on Mt Mavrovouni  
758 occurred during the last glacial cycle and most probably during the LGM. Moreover they are

759 well supported by local and regional morphostatigraphical correlations and are consistent  
760 with  $^{36}\text{Cl}$  surface exposure ages from limestone-derived moraines on the nearby Mt Tymphi  
761 as well as the Voidomatis River record within the same massif and the Ioannina basin pollen  
762 record. This study provides confidence in the suitability of  $^{36}\text{Cl}$  dating for iron-rich ophiolites  
763 using improved production rates from iron spallation, meaning that future studies can provide  
764 additional chronology for the highest peaks of the Pindus and the wider Balkans which are  
765 formed of a similar lithology.

766 The presented preliminary chronology was based on the geomorphological study, on the  
767 obtained ages as well as on the established morphostratigraphic correlations. At least three  
768 distinct Pleistocene cold phases were identified. Glaciers reached their local Late Pleistocene  
769 maximum within the global LGM interval at  $27.0 \pm 6.5$  ka. Cold conditions persisted until  
770  $20.2 \pm 4.8$  ka when a combination of retreating glaciers, permanent snow fields, permafrost  
771 and increased debris supply due to freeze-thaw weathering of the cirque walls led to the  
772 formation of periglacial landforms such as rock glaciers and pronival ramparts. This evidence  
773 suggests unfavourable (cold and arid) conditions for glacier development during this  
774 transitional phase from the global LGM to the Late-glacial but cold/humid enough to  
775 preserve perennial snow fields (i.e. névé fields) and small glaciers in topographically  
776 favourable positions supporting the hypothesis of increasing aridity prevailing across Western  
777 Balkans after the global LGM. The geomorphological and sedimentological evidence of a  
778 more extensive, pre-LGM glaciation phase(s) is also evident. This can be attributed to the  
779 Middle Pleistocene (MIS 12/MIS 6) on the basis of regional morphostratigraphic correlations.

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