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The glacial history of Mount Chelmos, Peloponnesus, Greece

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Abstract

Mount Chelmos in the Peloponnesus was glaciated by a plateau ice field during the most extensive Pleistocene glaciation. Valley glaciers radiated out from an ice field centred over the central plateau of the massif. The largest glaciations are likely to be Middle Pleistocene in age. Smaller valley and cirque glaciers formed later and boulders on the moraines of these glacial phases have been dated using ³⁶Cl terrestrial cosmogenic nuclide exposure dating. These ages indicate a Late Pleistocene age with glacier advance/stabilisation at 40-30 ka, glacier retreat at 23-21 ka and advance/stabilisation at 13-10 ka. This indicates that the glacier maximum of the last cold stage occurred during Marine Isotope Stage (MIS) 3, several thousand years before the global Last Glacial Maximum (LGM; MIS 2). The last phase of moraine building occurred at the end of the Pleistocene, possibly during the Younger Dryas.

Introduction

The mountains of Greece were glaciated on multiple occasions during the Middle and Late Pleistocene (Woodward and Hughes, 2011). Glaciated terrains have been identified on Mount Olympus in the east (Smith *et al.*, 1997) and throughout the Pindus Mountains (e.g. Sestini, 1933; Mistardis, 1952; Pechoux, 1970; Messerli, 1967; Hagedorn, 1969; Palmentola *et al.*, 1990; Mastronuzzi *et al.*, 1994; Boenzi and Palmentola, 1997; Woodward *et al.*, 2004; Woodward and Hughes, 2011). On Mount Tymphi in NW Greece, different generations of moraines have been dated and assigned to at least three separate cold stages: the Skamnellian Stage (Marine Isotope Stage [MIS] 12); the Vlasian Stage (MIS 6), and; the Tymphan Stage (MIS 5d-2) (Hughes *et al.*, 2006a).

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38 Moraines have been identified in mountains as far south as the Peloponnesus (Hagedorn, 1969;
39 Mastronuzzi *et al.*, 1994; Pope, 2013; Bathrellos *et al.* 2014) and also Crete (Fabre and Maire 1983;
40 Bathrellos *et al.*, 2014). In the Peloponnesus several different mountain areas display evidence of
41 extensive glaciation (e.g. Taygetos, Chelmos/Aroania, Ziria/Kyllini, and Erimanthos). However, until
42 now the geochronology and palaeoclimatic significance of these glacial features has remained
43 unexplored. The mountains of the Peloponnesus are strategically important for understanding the
44 relationship between glaciers and climate because of their position at the tip of the Balkan Peninsula.
45 It is well established that glaciers have a close relationship with climate (e.g. Ohmura *et al.*, 1992;
46 Hughes and Braithwaite, 2008) and temporally-constrained glacier-climate reconstructions from the
47 Peloponnesus will provide interesting comparisons with not only other glacial records in the
48 Mediterranean region (e.g. Hughes *et al.*, 2006b), but also with other records of environmental
49 change.

50

51 Most glacial records in Greece and the western Balkans have been dated using U-series (Hughes *et al.*
52 2006a, 2010, 2011a) which provides a minimum-age for the glacial deposits by dating secondary
53 carbonates. These secondary carbonates can form sometime after the formation of the host moraines
54 and this limits the precision of this particular technique with respect to obtaining the age of the host
55 moraine. Unlike U-series dating, cosmogenic nuclide analyses have the potential to obtain a direct age
56 for a moraine surface and thus provide a more precise geochronology for the Late Pleistocene
57 glaciations of Greece, as has been achieved in neighbouring Turkey (Akçar *et al.* 2016; Çiner *et al.*
58 2016; Sarıkaya *et al.*, 2016). On Mount Olympus in northeastern Greece, Manz (1998) was one of the
59 first to date glacial surfaces in the Mediterranean mountains using ^{36}Cl exposure dating. However,
60 there is considerable scatter in this dataset and interpreting the significance of these pioneering
61 exposure ages is difficult and this issue is discussed in Woodward *et al.* (2004) and Woodward and
62 Hughes (2011).

63

64 The timing of glacier advances during the last glacial cycle shows significant variation around the
65 world. Hughes *et al.* (2013) showed that many glaciers reached their maximum extents not at the
66 global last glacial maximum (LGM; 27-23 ka, Hughes and Gibbard 2015) in MIS 2 but earlier, during
67 MIS 3, 4 and even 5. In the Mediterranean mountains recent papers have revealed a similar story. In
68 the Pyrenees the largest glaciers of the Late Pleistocene appear to date from early in the last glacial
69 cycle, possibly as early as MIS 5 (Pallàs *et al.*, 2010; Delmas *et al.*, 2011). Evidence from the
70 Cantabrian Mountains of northern Spain suggests that glaciers reached their maximum Late
71 Pleistocene extent during MIS 3 or earlier (Jiménez-Sánchez and Farias Arquer, 2002; Moreno *et al.*,
72 2010; Serrano *et al.*, 2012; 2016; Nieuwendam *et al.* 2016) whereas evidence from other areas further
73 south, such as the Sanabria, Sierra de Gredos and Serra de Guadarrama, suggests that the most

significant advance occurred during MIS 2 close to the global LGM (e.g. Palacios et al. 2011; 2012; Rodríguez-Rodríguez et al. 2014). In Corsica, the LGM is clearly represented as the largest advance of the last cold stage (Kuhlemann et al., 2008) and so too directly to the north in the Maritime Alps of Italy (Federici et al., 2012). A similar pattern is also found in the Italian Apennines, although here the local glacier maximum occurred slightly earlier than the global LGM, between 32 and 27 ka (Giraudi 2012). In northern Greece, a similar age was given for the maximum extent of Late Pleistocene glaciers based on correlations between dated fluvial deposits and upstream glaciation (Hughes et al., 2006c). In Turkey, there is also now evidence for a pre-LGM advance with glaciers reaching their maximum positions by 35.1 ± 2.5 ka on Mount Akdağ in the southern part of the Taurus range (Sarıkaya et al., 2014).

In many places the highest cirque moraines are assumed to be of Late-glacial age. In the Italian Maritime Alps this has been confirmed using cosmogenic exposure dating and here there is also evidence for a glacier advance during Heinrich Event I as well as the Younger Dryas (Federici et al., 2012; 2008). Elsewhere the timing of Late-glacial advances remains speculative. For example, in northern Greece Hughes et al. (2006d) assumed that the highest un-dated moraines on Mount Smolikas, the highest of the Pindus, formed during the Younger Dryas. This is supported by evidence from Montenegro to the north. Here, U-series dating of secondary carbonate cements in moraines suggest that the last major moraine-building phase of the Late Pleistocene occurred during the Younger Dryas (Hughes and Woodward 2008; Hughes et al., 2010; 2011a). However, elsewhere in the eastern Mediterranean the evidence for Younger Dryas moraines has been elusive, although a recent papers by Çiner et al.(2015) and Sarıkaya and Çiner (2016) provide new evidence of Younger Dryas moraines in Turkey dated using ^{36}Cl exposure dating. In the eastern Balkans, in Kosovo, Younger Dryas moraines have also been dated using cosmogenics (^{10}Be) (Kuhlemann et al. 2009). Similar findings have also reported from as far east as Ukraine (Rinterknecht et al. 2012). In Greece, whilst the Younger Dryas is represented in cave sediments (Karkanis 2001), diatom records from lake sediments (Wilson et al. 2008) and alluvial fan sediments from the Sparta basin (Pope and Wilkinson, 2006), it is not so clearly represented in pollen records (Bottema 1995; Lawson et al. 2004). Questions therefore remain as to whether or not the Younger Dryas cold interval is represented by glaciers in the mountains of Greece, and also what the presence or absence of glaciers tells us about climate at this time.

The aim of this project is to establish the extent, timing and palaeoclimatic significance of Late Pleistocene glaciations in the Peloponnese. The focus is on dating the most recent deglacial history in the highest cirque areas and comparing the deglacial chronologies with other Mediterranean mountains. This paper uses ^{36}Cl analyses to provide exposure ages for the most recent series of moraine surfaces on Mount Chelmos (2355 m a.s.l.) in the northern Peloponnese, Greece (Fig. 1).

111 Glacier-climate reconstructions for different glacier phases are then compared with evidence from
112 other glaciated Mediterranean mountains in order to establish former climatic patterns for the last
113 glacial cycle. Mount Chelmos represents one of the southernmost glaciated areas of Europe (38°N),
114 rivalled only by the Sierra Nevada of Spain and Etna, Sicily (although in the latter area any glacial
115 traces have been obscured by volcanic activity). Thus, glacier records in the Peloponnesus provide a
116 unique insight into palaeoclimatic conditions in the Mediterranean, especially for the eastern
117 Mediterranean basin.

118

119 Study areas

120 Mount Chelmos (also known as Aroania) reaches 2355 m a.s.l. and is the third highest mountain in
121 the Peloponnesus. The massif forms part of the mountain chain of the External Hellenides and
122 consists of rocks belonging to three distinct tectonic units. The upper one is the Pindos Unit, which
123 occupies both the highest peaks and the southern slopes of the mountain and consists mainly of
124 pelagic limestones and radiolarites of Mesozoic age overlain by a Tertiary clastic sequence (flysch).
125 The Tripolis Unit is the intermediate unit and consists of platform carbonates that are Upper Triassic
126 to Upper Eocene in age overlying a volcanosedimentary sequence, called Tyros Beds. The lower
127 altitude slopes of northeastern Chelmos are built by HP-LT metamorphic rocks of the lowest
128 Phyllites-Quartzites Unit. These units were folded and stacked together during the main alpine
129 orogenesis culminated in the Oligocene and Miocene while subsequent syn-orogenic extension
130 resulted in exhumation of metamorphic rocks (Skourtsos and Kranis, 2009; Skourtsos and Lekkas,
131 2010). Although, plate movements and orogenesis are still active far away to the west, the Gulf of
132 Corinth, immediately to the north of Mount Chelmos, is a very active and fault-controlled rift with
133 contemporary uplift of the Peloponnesus leading to new faulting (Moretti et al, 2003; Skourtsos and
134 Kranis, 2009; Ford et al, 2013) (Fig. 1). The northern areas of the mountain forms the southern, now
135 inactive, margin of the rift at the hanging wall of which continental Pliocene deposits have been
136 deposited (Skourtsos and Kranis, 2009; Ford et al, 2013). Pleistocene and Holocene estimated uplift
137 rates in the northern Peloponnesus are not homogeneous and range from 0.3 mm yr⁻¹ 50-60 km
138 northeast of Mount Chelmos (Collier et al., 1992) to 1.5 mm yr⁻¹ (Armijo et al., 1996; McNeill et al.,
139 2004; Palyvos et al., 2008) and even higher, 2.9 to 3.5 mm yr⁻¹, during the Holocene (Pirazzoli, et al.,
140 2004) 20 km north of Mount Chelmos, and 0.55 mm yr⁻¹ in the southern Peloponnesus (Mariolakos et
141 al., 1994).

142

143 The nearest official climate normal data is available from Patrai (Patras) station (3 m a.s.l.) c. 50 km
144 NE of Mount Chelmos. The 1961-1990 period at this site recorded a mean annual temperature of
145 17.9°C, a mean annual range of 16.7°C and mean annual precipitation of 682 mm (World
146 Meteorological Organisation, 1998). The climate at Kalavryta weather station (731 m a.s.l.), at the

foot of Mount Chelmos, has recorded a mean annual temperature of 13.6°C and mean annual precipitation of 996 mm. In the highest mountain areas annual precipitation must therefore exceed 1000 mm and is estimated to be in the range 1500–2000 mm based on the precipitation difference between Patras (3 m a.s.l.) and Kalavryta (731 m a.s.l.). Precipitation is strongly seasonal with a wet winter season and an arid summer season, which is typical of Mediterranean climates (Koutsopoulos and Sarlis, 2003). No glaciers survive today on Mount Chelmos, although snow patches have been observed by the authors in late July below the north-facing cliffs near Mavrolimni (Fig. 2).

Methods

Field methods

The glacial geomorphology of Mount Chelmos was mapped in the field in July 2010 and July 2011. Place names are based on the 1:50,000 topographic map (Anavasi, 2008). The accuracy of this field-mapping was aided by high-resolution satellite imagery available in GoogleEarth. Moraine units were identified on the basis of sedimentological and morphological criteria (e.g. clast size, shape, roundness, sediment sorting, surface form) and subdivided on the basis of morphostratigraphy (Hughes, 2010).

Several studies have successfully dated limestone landslide boulders from Marine Isotope Stage (MIS) 2 and later using ^{36}Cl analysis (e.g. van Husen *et al.* 2007, Ivy-Ochs *et al.*, 2009) although there are few examples where older surfaces have been successfully dated using this technique. However, in other areas of Greece the oldest moraines have been shown to be Middle Pleistocene in age, as old as MIS 12 (Hughes *et al.* 2006a). Samples were therefore not taken from the lowest moraines on Chelmos because there is a much greater risk of age scatter from these moraines because of exhumation, toppling and erosion/weathering of the boulder surfaces. Instead, we targeted the highest moraines (and therefore stratigraphically youngest) in selected valleys and sampled seven glacial boulders using a hammer and chisel (Table 1).

All of the samples were from fine-grained crystalline limestones. All samples were taken from the tops of boulders at least 50 cm away from any edges in order to minimise edge effects which are most significant for terrestrial cosmogenic nuclides such as ^{36}Cl , which is produced in part through thermal neutron capture (Gosse and Phillips 2001).

Laboratory analysis and geochronological calculations

Rock samples were crushed and sieved. The $<63\ \mu\text{m}$ fraction was mixed with resin and pressed to form a powder pellet for XRF analysis of the bulk rock (major elements % and some trace elements ppm) s. Some trace elements (Li, B, Sm, Gd, Th, U) were also determined using ICP-MS for higher

precision (ppm). The 250-500 μm fractions were sent to PRIME Lab at Purdue University for ^{36}Cl processing and accelerator mass spectrometry (AMS) measurement.

The sample data and the results of the AMS and geochemical analyses from the rock samples (Tables 1 to 3) were used to calculate exposure ages using the excel calculator of Schimmelpfennig et al. (2009). This calculator has been used to provide ages for limestone samples as well as basalt samples. Examples of its successful use to date carbonate surfaces include Le Dortz et al. (2011), Sadier et al. (2012) and Ryb et al. (2014). The default production rate for the spallation of Ca in Schimmelpfennig et al. (2009) is from Stone et al. (1996) at 48.8 ± 3.4 atoms ^{36}Cl (g Ca) $^{-1}$ a $^{-1}$. Other values for the production rate for Ca are 42.8 ± 4.8 atoms ^{36}Cl (g Ca) $^{-1}$ a $^{-1}$ from Schimmelpfennig et al. (2011), 56.0 ± 2.2 atoms ^{36}Cl (g Ca) $^{-1}$ a $^{-1}$ from Marrero (2012) and 57 ± 5 atoms ^{36}Cl (g Ca) $^{-1}$ a $^{-1}$ from Licciardi et al. (2008). However, in all three of these papers the production rates are derived from mixed rock compositions (rocks with both Ca and K present) whereas Stone et al. (1996) is derived from limestones (with just Ca present). Thus, the production rate of Stone et al. (1996) is likely to be the most reliable for limestone samples.

Bulk rock densities were measured in the laboratory using a simple water displacement method and were c. 2.5 to 2.7 g cm $^{-3}$, which is consistent with most limestones. Scaling factors for the production of nucleonic and muonic production as a function of elevation, latitude and temporal variations were calculated following Stone (2000). Correction factors for shielding of a sample of arbitrary orientation by surrounding topography were calculated in the CRONUS on-line calculator (*Version 1.1. March, 2006. Written by Greg Balco; http://hess.ess.washington.edu/math/general/skyline_input.php*). All samples were taken away from edges from the top surface of boulders that were at least 1 m high. Consequently, correction factors for geometry effects on spallogenic production and for snow shielding for spallogenic production were not necessary and given a value of 1 in the excel calculator. A value of 177g cm $^{-2}$ was used for the effective fast neutron attenuation coefficient. The estimated ^{36}Cl concentration from inheritance was assumed to be zero. The measured ^{36}Cl concentration in the sample was reported by PRIME Lab. Inputs for exposure duration, formation age of rock for radiogenic correction, and erosion rate were left blank in the calculator as we assume that surfaces have undergone minimal surface loss since glaciation.

Glacier reconstructions

The combination of the geomorphological mapping and the ^{36}Cl exposure ages will provide the basis for temporally-constrained glacier-climate reconstructions. Former glaciers were reconstructed based on limits defined by moraine crests (where present) or the distal limits of glacial deposits. The upper limits of the former glaciers were constrained by cirque morphology and the altitude of the cirque

headwall. In the absence of a cirque then other criteria were used to constrain upper ice limits, such as the upper distribution of ice-moulded bedrock, perched boulders and periglacial trimlines. Glacier surface profiles for the most extensive ice cap phase were reconstructed using an iterative flowline model (Benn and Hulton, 2010). This approach takes into account topographical irregularity below the former ice caps. A basal shear stress of 50-100 kPa was applied based on a range of measurements known to occur in valley glaciers (see Vieira 2008, p. 197 for further references). This produced a range of potential surface altitudes for the former ice masses and the most suitable end of this range (50 or 100 kPa) was constrained using geomorphological evidence of glacial erosion and deposition at certain altitudes. For the smaller cirque and valley glacier phases glaciers were simply reconstructed using the geomorphological evidence (moraine positions primarily) and tested for plausibility by calculating the basal shear stress. The shape and distribution of surface contours and ice margins (where not constrained by geomorphology) were then tuned until shear stress values in the range 50-100 kPa were obtained.

Equilibrium line altitudes (ELAs) are reported in this paper using an accumulation area ratio (AAR) of 0.6. This is because this value was used to reconstruct the ELAs on the glaciers in NW Greece on Mount Tymphi and Mount Smolikias (Hughes et al. 2006c; 2006d; 2007a). However, in order to make comparisons with glaciers around the Mediterranean, a range of potential ELAs were also estimated using a range of AARs from 0 to 1. For example, the ELAs of different generations of glaciers on Mount Orjen, Montenegro, were calculated using AARs of 0.5, 0.6 and 0.8 (Hughes et al 2010). These examples were based on the AAR with the lowest standard deviation in a population of glaciers of the same age (cf. Osmaston, 2002; Hughes et al., 2010). This approach assumes that for a sample of similar types of glaciers on the same mountain, the controls on ELA position will be similar, producing comparable patterns in glacier hypsometry. Often, glacier snouts may be expected to have the greatest variation in altitude (with an AAR of 1) because glacier front positions depend on the overall hypsometry of the glacier above it. Conversely, glacier sources (with an AAR of 0) are often less variable on a single radially-glaciated mountain, such as Chelmos, because peak altitudes across the glacier sample tend to be similar. This relationship between snout and source altitudes across a glacier sample tends to produce an asymmetrical U-shaped relationship for populations of valley and cirque glaciers (Osmaston, 2000, Fig. 9.4.7, p. 182). The standard deviation of AAR approach is simply a tool that can help describe the former hypsometry of glaciers and a variety of patterns may emerge depending on the geometry of the former glaciers. Different AARs may be suitable for different types of glaciers, and it is known that the ELAs of ice caps, valley glaciers and cirque glaciers correspond to different AARs (e.g. Leonard 1984) as do debris-covered versus clean glaciers (Clark et al., 1994). Ultimately, the user must still rely on empirical observations of the relationship between AARs and ELAs and be aware of potentially wide variability for palaeo glaciers.

255 It is known that Mount Chelmos is situated in one of the most tectonically active parts of Greece, just
256 south of the Gulf of Corinth. The potential for tectonic movements is therefore great. Comparing
257 ELAs with other parts of Greece (e.g. Hughes et al. 2006c; 2007a) is problematic given the
258 uncertainties in knowing the uplift history of the area. This issue is explored further in the Discussion.
259

260 Results

261 *Glacial geomorphology*

262 The extent and distribution of glacial landforms on Mount Chelmos is illustrated in Fig. 2. The
263 characteristics of the evidence for glaciation are described below for the five main valleys that drain
264 the massif.
265

266 (1) Spanolakos Valley

267 The Spanolakos valley drains the northwestern flank of Chelmos. A series of sediment landforms are
268 present in this valley and these are described then interpreted in morphostratigraphical order from
269 lower to upper valley.
270

271 *Unit 1*

272

273 *Description:*

274 The lower slopes, below c. 1600-1700 m are characterised by thick accumulations of thick
275 accumulations of gravels separated by occasional sandy interbeds. These are exposed by the modern
276 river channel and reach thicknesses of >30 m. The sediment spreads out from the apex (at c. 1600-
277 1700 m) to reach a width of nearly 2 km in the lower parts at c. 1200-1300 m near the village of Kato
278 Lousi. Between 1600-1700 m cemented gravels and sands are topped by numerous large perched
279 boulders (Fig. 3). These have diameters of >1 m and are subangular to subrounded.
280

281 *Interpretation:*

282 The large area of bedded gravels and sands that reach down to Kato Lousi is interpreted as a stacked
283 sequence of glaciofluvial fans. The perched boulders in the upper part of the fan are interpreted as
284 glacial in origin and are similar to deposits found in several other valleys (see below). The boulders
285 do not form clear moraines but are instead scattered over wide areas. The down-valley limit of these
286 boulders is mapped as a boulder limit in Fig. 2 and is interpreted as the approximate down-valley
287 extent of the largest former glacier in this valley. These glacial deposits are denoted as stratigraphical
288 unit 1 in Fig. 2.
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290 *Unit 2*

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Description:

At c. 1900-2050 m a very clear arcuate ridge is present (Fig. 4). Subrounded and subangular boulders (many >1 m in diameter) are present on the surface of this feature. The ridge is exposed near the modern river channel and this reveals >10 m thickness of diamicton, characterised by subrounded/subangular boulders supported by a pale sandy-silt matrix (clast density: c. 20%). The sediments are cemented in several places and cemented horizons are present.

Interpretation:

The sediment ridge is mapped as a moraine ridge marking the termino-lateral positions of a former valley glacier. This was also the interpretation of Mastronuzzi et al. (1994, their Fig. 4). These glacial deposits are denoted as stratigraphical unit 2 in Fig. 2.

Description:

Between c. 1900-2000 m the valley floor is filled with boulders, some of which reach several metres in diameter. This sediment pile forms a subdued ridge that bounds the northern side of the main river for c. 250 m with a width of c. 75 m Subrounded and subangular boulders (many >1 m in diameter) are present on the surface of this feature .

Interpretation:

The sediment ridge is mapped as a moraine/glacial boulders (indistinct) in Fig.2 marking the terminus of a former cirque glacier. This small glacier would have been banked up against the unnamed peak with spot height 2145 m whose northern cliffs mark the headwall of this cirque. These glacial deposits are denoted as stratigraphical unit 3 in Fig. 2.

(2) Laghada valley (Strogilolaka & Kato Kambos)

The Laghada valley also drains the western slopes of Chelmos and is the southern neighbouring valley of the Spanolakos (Fig. 2). However, the Laghada valley is very different in profile to the Spanolakos. The Laghada has a very steep middle profile (from 1400-2000 m) and forms a deeply incised valley for the entire middle and lower reaches (from 1100-2000). The upper valley (above 2000 m) is perched with a gentle gradient. This valley morphometry means that the preservation potential for Pleistocene deposits in the middle and lower Laghada is very limited. All of the evidence for glaciation is restricted to the highest reaches where the Laghada drains two elevated pastures near the shepherd pastures at Strogilolaka and Kato Kambos (Fig. 2).

Unit 1

328 *Description:*

329 At Strogilolaka large subrounded and subangular boulders are scattered over wide areas between
330 altitudes 1900 and 2100 m immediately below and west of the bulldozed track that traverses the
331 western flank of Chelmos. In several places these lie on top of diamicton ridges (boulders in a silt
332 matrix) that are several metres high. These diamicton deposits have been heavily incised by modern
333 stream activity and are poorly preserved. The eastern side of this area is bounded by steep rocky
334 slopes until the crest of the watershed with angular debris present.

335 In the southern fork of the Strogilolaka valley leading up towards Kato Kambos occasional boulders
336 can be seen perched in the steep valley. However, the river has a very steep course, which limits
337 preservation of unconsolidated deposits. In addition access is difficult to this area and observations
338 were limited.

339

340 *Interpretation:*

341 These are interpreted as glacial deposits and correlated with the Unit 1 deposits in the Spanokakkos
342 valley. No other glacial deposits are present up-valley of this area, only talus and rock slopes failure
343 deposits which form some prominent ridges of large angular boulders below the west-facing cliffs
344 which are traversed by the bulldozed track immediately up-valley of the Strogilolaka moraines.

345

346 A glacial limit is also tentatively mapped in the southern fork of the Strogilolaka valley which lead up
347 towards Kato Kambos, although moraines are not clearly preserved.

348

349 *Unit 2*

350

351 *Description:*

352 A sediment ridge with large subrounded and subangular boulders on its surface is situated on the
353 northern side of the valley immediately west of Kato Kambos (Fig. 2) between c. 2050 and 2140 m.

354 The boulders have diameters of 1-3 m and many are present along the ridge crest. Boulders are also
355 present on the south-side of the modern river channel (again between c. 2050 and 2140 m) although
356 these are isolated and scattered and do not form a discernible ridge crest.

357

358 *Interpretation:*

359 The sediment ridge in the lower Kato Kambos clearly resembles a moraine ridge and is interpreted as
360 a latero-frontal moraine of a former valley glacier. The lower limit of boulders in this area on both the
361 north and south side of the modern river channel suggests a down-valley altitude of c. 2050 m for this
362 glacier, perched just above the steep and deeply incised Lagadha valley. These moraines and glacial
363 boulders are correlated with unit 2 in other valleys on the basis of altitudinal position.

364

Two boulders from the outermost latero-frontal moraine were sampled for ^{36}Cl analysis, both from the crest at c. 2120 m altitude. The boulders are shown in Fig. 5 with sample details and exposure age results shown in Tables 1, 2 and 3. These samples yielded exposure ages 39.6 ± 3.0 and 30.4 ± 2.2 ka.

Unit 3

Description:

A very clear arcuate sediment ridge is present with a crest situated at c. 2140 m with numerous large (>2-3 m) subrounded boulders scattered on the surface is present bounding the basin of Kato Kambos which is situated below the northern cliffs of Profitis Ilias (2282 m). The ridge crest is > 10 m above the floor of the basin and almost completely closes the NW outlet of the basin with only a small stream cutting through the ridge. The floor of the Kato Kambos basin contains a level area of marshland. A subdued sediment ridge with surface boulders is also present in the eastern part of the basin in a small adjacent hollow (Fig. 2).

Interpretation:

Kato Kambos is a glaciokarst cirque basin. The backwall cliffs of Profitas Ilias provided the sources of the former glacier. The basin is over-deepened and is similar in appearance to glaciokarst basins observed in northern Greece (Hughes et al. 2007b) and Montenegro (Hughes et al. 2011a). The cirque at Kato Kambos (Fig. 6) is one of the best-developed cirques on Chelmos, alongside the cirques at the head of the Chaliki valley below spot height 2318 m, at the head of the Xerokambos valley, and below spot height 2145 m at the head of the Spanolakkos valley; in other areas cirques are poorly-defined or not present at all. The sediment ridge is interpreted as an arcuate moraine and represents one of the best and clearest examples of a cirque moraine in Greece. The marshland in the centre of the cirque basin hints at the presence of potential lake and a break of slope encircling the basin (c. 5-10 m above the basin floor) may indicate a former lake shoreline. A subdued moraine crest is also present in the eastern basin (Fig. 2) and this may be associated with glacier retreat as the cirque glacier would have split into two cirque basins. These moraines represent the highest moraines in the Strogilakkos valley and are correlated with Unit 3 moraines in other valleys.

A large boulder (diameter >3 m, height > 2m) on the crest of the moraine was sampled for ^{36}Cl analysis (Fig. 7). The sample details and exposure age results are shown in Table 1, 2 and 3. This sample yielded an exposure age of 12.6 ± 0.9 ka (CH10). Another large boulder (diameter >2 m, height >1 m), just 5 metres distance from CH10 was also sampled and this yielded an exposure age of 10.2 ± 0.7 ka.

(3) Chaliki Valley

402 The Chaliki valley drains the southern flank of Chelmos immediately south of the steep southern rock
403 face of Psili Korfi (Fig. 2). The lower valley is characterised by large spreads of gravel and sands and
404 a braided river system leading to alluvial terraces near the village of Planitero. Glacial deposits are
405 restricted to the higher reaches and these are described below.

406

407 *Unit 1*

408

409 *Description:*

410 Diamicton deposits with large subrounded boulders within a silt matrix are exposed in the modern
411 river channel at an altitude of c. 1300 m (Fig. 8) immediately to the north of Magherou, a southeastern
412 spur of Profitis Ilias. The valley is very steep and narrow and these deposits are found plastered
413 against the valley sides, often well above the modern river channel but have little or no surface form.
414 Some boulders display evidence of striae.

415

416 *Interpretation:*

417 The diamicton deposits are interpreted as glacial in origin, formed by a valley glacier that extended
418 down this valley. The poorly sorted materials and presence of a fine silt matrix along with striated
419 boulder clasts is the main basis of this interpretation. The lowermost presence of these deposits is
420 taken as an approximate limit of glaciation. However, the steepness of the catchment means that
421 preservation potential is low. These glacial deposits are correlated with stratigraphical unit 1.

422

423 *Unit 2*

424

425 *Description:*

426 Mounds of diamicton deposits with large (>1 m) subrounded and subangular boulders in a silt matrix
427 are situated above a break in slope of the main river channel between altitudes of c. 1550-1650 m. The
428 valley becomes more open in this area and the river bed widens considerably forming a wide braid
429 plain. The diamicton deposits are exposed by numerous smaller tributary streams in this area and a
430 concentration of scattered and perched subrounded and subangular boulders occurs throughout this
431 area. Similar boulder diamictons are also exposed in the centre of the valley on an interfluvium
432 separating two stream channels between 1680 and 1900 m (see Fig. 2).

433

434 *Interpretation:*

435 The concentration of diamicton deposits and perched boulders in this area are interpreted as eroded
436 end moraines. These moraines are correlated with Unit 2 moraines found in other valleys.

437

438 *Unit 3*

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Description:

Clear sediment ridges are present in the basin below the SE face of the summit spot height 2318 m marked on topographic maps (Anavasi, 2008) and bound an area of concentrated boulder debris (Fig. 9). The sediments ridges reach an amplitude of nearly 10 m whilst boulders within these limits reach diameters of >5 m and are an admixture of angular, subangular and subrounded boulders.

Interpretation:

The sediment ridges are interpreted as the end moraines of a small cirque glacier. The concentrated boulder debris inside of these moraines is a mixture of subglacial and supraglacial debris as well as talus debris. Whilst the cirque is clear as an armchair-shaped bowl, it is weakly incised (Fig. 9).

(4) Neraidhorachi

The Neraidhorachi area represents the central plateau area of Chelmos and a major valley drains northeastwards towards the village of Mesorroughi. The valley is the deepest and widest of all the valleys draining Chelmos. The Styx waterfall, named after the river in ancient Greek mythology and today part of the Mavroneri River, occurs in the upper valley area below the Neraidhorachi plateau (Mayor and Hayes 2011).

Unit 1

Description:

Diamicton deposits and spreads of perched subrounded and subangular boulders are present in the forests and reach below an altitude of 1200 m. These deposits form extended ridges on both side of the modern river valley that drains towards Mesorroughi (Fig. 2). The sediment ridge is most clearly expressed between c. 1180-1450 m on the northern side and is largely covered in forest. However, surface boulders are clearly visible in small clearings and revealed in section by forest roads.

Interpretation:

The diamictons and boulders have the characteristics of glacial deposits in that they are poorly sorted, subrounded and subangular and form ridges on both sides of the valley. The sediment ridges resemble latero-frontal moraines and are consistent with a large outlet glacier draining from the central plateau area of Neraidhorachi. The valley morphology in the area of the Styx waterfall also resembles an over-steepened U-shaped valley indicating that a glacier must have extended down-valley far beyond this locality.

Unit 2

476

477 *Description:*

478 Mounds of subrounded boulder-rich diamicton are present at several positions in the valley northeast
479 of Neraidhorachi and are best preserved where the long profile of the valley becomes less steep, such
480 as at c. 1600, 1800, 2000, 2100 and 2250 m. These successive piles of diamicton and boulder material
481 are largely shapeless, although form irregular mounds and occasionally ridge-like forms, the latter
482 especially so in the higher areas. For example, around Epano Kambos, an arcuate sediment ridge
483 encloses the uppermost basin at c. 2250 m.

484

485 *Interpretation:*

486 The diamicton mounds and boulder accumulations are interpreted as moraines of a former valley
487 glacier. The moraine unit spans the valley from 2250-1600 m and is separated from Unit 1 because of
488 the large distance and altitude between these and the lowermost glacial deposits in this valley, which
489 are 2 km and more than 400 m in altitude down-valley. The multiple moraines, situated at the breaks
490 in the valley profile, are consistent with terminal and recessional positions of the former glacier that
491 occupied this valley. The moraines are grouped together morphostratigraphically with the lower limit
492 of these deposits at 1600 m taken as the lower limit of the glacier. The glacier occupied a large U-
493 shaped valley that was inherited from the larger glacier associated with the Unit 1 moraines 2 km
494 down-valley. The successive moraine piles at 1800, 2000, 2100 and 2250m are interpreted as
495 recessional or stand-still moraines as the glacier retreated up-valley. These moraines are formed of
496 substantial accumulations of sediment (>10 m thickness) and are incised by the modern stream
497 channel. The uppermost arcuate moraine at Epano Kambos is very clearly defined and could be
498 classified as a separate morphostratigraphical unit and correlated with Unit 3 moraines in other
499 valleys. However, the moraine has a similar surface exposure age to lower moraines in this area – see
500 below and Fig. 2. Furthermore, the basin headwall is less steep than in other areas occupied by Unit 3
501 moraines elsewhere where a steep cirque backwall always exists. The basin in Epano Kambos, whilst
502 cirque-like, lacks a steep backwall, and is therefore not mapped as a cirque in Fig. 2 despite the
503 presence of a moraine.

504

505 Three boulders from the uppermost moraine mounds (>2200 m a.s.l.) in this assemblage were
506 sampled for ³⁶Cl analysis, two from the crest of a broad moraine ridge on the west side of the valley
507 and another from the crest of the uppermost arcuate moraine. The sample details and exposure age
508 results shown in Tables 1, 2 and 3. These samples yielded exposure ages of 21.2 ± 1.6 , 21.6 ± 1.6 and
509 22.9 ± 1.6 ka, respectively. These ages are indistinguishable and overlap within error and indicate that
510 ice must have retreated rapidly at 23-21 ka in the highest parts of this valley (>2200 m a.s.l.). The
511 lowest moraines of this unit, which were associated with a glacier that reached down to 1600 m a.s.l.
512 must pre-date the exposure ages from the uppermost moraines.

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(5) Xerokambos

The Xerokambos valley drains northwards from Chelmos. The valley contains the Kalavryta ski centre and is the easiest valley to access. There has been considerable reworking of surface deposits in this valley due to construction activities.

Unit 1

Description:

Scattered subrounded boulders are present throughout the area of Xerokambos. A thin layer of diamicton with cobbles and boulders lodged within the deposits is present also. The Ski Centre and adjacent car park has destroyed much of the evidence in the lower valley area.

Interpretation:

Mastronuzzi et al. (1994, p. 81) recognised that “a veneer of morainic deposits” occurs in the valley floor down as far as the ski centre. The observations made above of large subrounded boulders and thin diamicton deposits supports their interpretation that a glacier once reached as far as the Ski Centre buildings and car park. There is no clear moraine limit since much of the ground has been disturbed but this area is likely to define the lower limit of the most extensive former glacier that has existed in this valley.

Unit 2

Description:

A well-preserved arcuate diamicton ridge containing numerous subrounded boulders is present at c. 1800 m altitude. The arcuate ridge is c. 50 m high and 500 m wide. Bedrock up-valley of the ridge is smoothed and the valley headwall is distinctly concave.

Interpretation:

The arcuate ridge of boulder diamicton is interpreted as an end moraine of a former cirque glacier that emanated from the northwestern cirque of the highest peak of Chelmos. This clear glacial feature was also recognised by Mastronuzzi et al. (1994). The smoothed bedrock up-valley of this ridge was interpreted as roche moutonnées by these authors and the field evidence does suggest ice moulding, although there has been considerable reshaping of the floor of this cirque in recent years in developing ski runs.

550 Unit 3

551

552 *Description:*

553 A small accumulation of boulders occurs on the west side of the uppermost cirque at c. 2180 m
554 altitude. The evidence from the centre of the valley has been obliterated by bulldozing of a ski run.
555 The uppermost cirque is very enclosed, much more so than the lower parts of the valley.

556

557 *Interpretation:*

558 The tight cirque basin at the head of Xerokampos is likely to have hosted a small cirque glacier. The
559 small accumulation of boulders on the west side of the valley, which are interpreted as moraines,
560 supports this hypothesis. This moraine unit is correlated with Unit 3 moraines, the highest present in
561 other valleys.

562

563 (6) Watersheds and interfluves

564 This section describes the geomorphology of the highest plateaus and ridges that separate the various
565 valleys described above. This is important for determining whether glaciers were restricted to their
566 valleys or submerged the landscape in the form of an ice cap or plateau ice field.

567

568 *Description:*

569 Large areas of bare smooth bedrock occur to the north of Psili Korfi including the entire western flank
570 of spot height 2253 m (Fig. 2). Large subrounded and sub angular boulders are perched on some
571 interfluves, including one very large boulder (>2 m diameter) just south of the summit of spot height
572 2253. Perched boulders are also present on the ridge col west of Psili Korfi summit. Bare bedrock
573 knolls are also present in the Neraidhorachi area. A prominent knoll occurs close to a shepherd
574 settlement in Epano Kambos, just east of the track leading to the astronomical observatory.

575

576 *Interpretation:*

577 The large areas of bare smooth bedrock are interpreted as ice moulded bedrock. The large subrounded
578 and subangular boulders that are found perched on interfluves are interpreted as glacially-transported
579 boulders. The source of these boulders is likely to be the ridge between Psili Korfi and spot height
580 2253 m. These observations have important implications for the configuration of ice masses on
581 Chelmos, since it means that ice completely submerged the central plateau area covering and eroding
582 interfluves and also transporting boulders at high levels. However, the ice mass was constrained and
583 hemmed in by the highest summits that surround the plateau.

584

585 *Summary of the geomorphological evidence*

586

587

588 The valleys of Chelmos all display clear evidence of multiple glacial phases in the form of moraines.
589 There are three morphostratigraphical glacial (moraine) units in the valleys and cirques of Chelmos
590 (Table 4). Each of these morphostratigraphical units represents suites of moraines. In other words, the
591 moraine units are diachronous with outer moraines and inner (recessional) moraines grouped as part
592 of the same stratigraphical unit. The upper valley source areas are not always defined by cirques. The
593 clearest cirques are present at the heads of the Laghada valley (at Kato Kambos), the Chaliki
594 valley, the northeastern valley draining Neraidhorachi (at Epano Kambos) and at the head of the valley
595 above Xerokambos (above the ski centre) (see Fig. 2). There is also evidence of ice moulding and
596 boulder transport at high-elevations close to watersheds and this indicates that ice must have
597 submerged the central plateau areas of Chelmos with an ice cap centred over the Neraidhorachi area
598 which marks the central plateau of Chelmos. This explains the apparent paradox that there are only
599 four clear cirques yet extensive moraine successions in six valleys that radiate from the centre of the
600 massif (see Fig. 2 and Fig. 10). The geomorphology is therefore crucial for the correct reconstruction
601 of former ice masses corresponding with the different morphostratigraphical moraine units. This is the
602 focus of the next section.

603

604 *Glacier reconstructions*

605 The largest, Phase I, glaciation was characterised by a contiguous ice cap covering an area of 9.74
606 km² that was radially drained by six outlet valley glaciers (Fig. 10). The longest outlet glaciers were
607 the Chaliki and Neraidhorachi glaciers which reached c. 3.5 km in length. The shortest outlet glacier
608 was the Strogilolaka glacier which had a length of c. 1.5 km. Ice thickness reached c. 500 m in the
609 Neraidhorachi valley north and northeast of Mavrolimni (Figs 1 and 2). The average ELA (assuming
610 an AAR of 0.6) of the six outlet glaciers was 1984 m a.s.l. There was no asymmetrical U-shaped
611 relationship between the standard deviation of altitude and AAR (e.g. Osmaston 2002). Instead there
612 was a positive relationship between AAR and standard deviation of altitude with the greatest standard
613 deviations for higher AARs (Table 5).

614

615 The Phase II glaciation was characterised by cirque and valley glaciers. A transect glacier covered the
616 watershed between Profitas Ilias and Psili Korfi with valley glaciers descending east and west. The
617 Phase II glaciers covered an area of 4.445 km². The smallest glacier was the Spanolakos glacier
618 (0.365 km²) whilst the largest was the Neraidhorachi (1.623 km²). The average ELA of the five
619 glaciers using an AAR of 0.6 was 1986 m a.s.l. The apparent paradoxical situation of much smaller
620 glaciers than Phase I but with similar ELAs is because of effects of the hypsometry of the Phase I ice
621 field in increasing the area of ice at higher elevations. This is examined further in the discussion.

Again, there was no asymmetrical U-shaped relationship between the standard deviation of altitude and AAR and the pattern was similar (but with smaller standard deviations) to the Phase I glaciers (Table 5).

The Phase III glaciers were small cirque glaciers. The four glaciers covered a total area of just 0.732 km², with the largest glacier in the Kato Kambos cirques (0.417 km²) and the smallest in the Xerokambos cirque (0.050 km²). These glaciers had an average ELA of 2114 m a.s.l. when using an AAR of 0.6. Again, there was positive relationship between the standard deviation of altitude and AARs of 0.4 to 0.8 (Table 5). In this sample of glaciers there was little variation in standard deviations across AARs.

Discussion

Mount Chelmos displays clear evidence of glaciation. Moraines are present down to 1200 m a.s.l. in five valleys draining radially from the central plateau area. The most extensive glaciation was characterised by an ice cap which submerged the highest areas with ice radiating outwards from a central ice dome located in the Neraidhorachi area. The resulting effects of ice cap glaciation are exhibited by evidence of ice moulding and glacier transport at high elevations, a radial glacial valley system. The largest and deepest glaciated valley was formed by a glacier that extended c. 4 km northeastwards from the central plateau (the length of the former glacier from the ice cap centre was c. 3.5 km).

The largest ice cap glaciation remains undated. Based on correlations with northern Greece and Montenegro (Hughes et al. 2006a; 2010; 2011a) it is likely that the largest glaciation is Middle Pleistocene in age (Skamnellian Stage/MIS 12 and Vlasian Stage/MIS 6). On Mount Chelmos the largest glacial phase is represented by a single morphostratigraphical unit (stratigraphical unit 1 in Fig. 1). However, it is possible that more than one glaciation is represented by this unit. At present it is not possible to separate out multiple moraine units that could correspond to both Skamnellian and Vlasian Stage glaciations.

Moraines at the heads of valleys on Chelmos are Late Pleistocene in age. The ³⁶Cl exposure ages from moraines at Kato Kambos (40-30 ka) suggest that the Late Pleistocene glacier maximum predates the global LGM. In other valleys the maximum Late Pleistocene moraines (associated with Unit 2 in Fig. 2) are undated. Nevertheless, whilst the succession of lower moraines ascribed to Unit 2 in the Neraidhorrissi valley (at c. 1600 m a.s.l.) are undated higher more proximal moraines (>2200 m a.s.l.) are dated to 23-21 ka. Thus, it is likely that moraine units depicted in Fig. 2 are diachronous reflecting maximum extents and later retreat phases.

658

659 The largest Late Pleistocene glacier emanated northeastwards from the Neraidhorachi area and
660 reached a length of c. 2 km. In this valley glacier retreat appears to have been rapid with two
661 successive moraine ridges in the highest part of the valley yielding ^{36}Cl ages that are indistinguishable
662 within error with the last glacier in this valley retreating at 23-21 ka, just after the global LGM which
663 occurred between 27 and 23 ka (Hughes and Gibbard 2015). Glacier retreat at 23-21 ka (± 1.6 ka) is
664 consistent with the evidence from northern Greece of dry conditions around the time of the LGM (cf.
665 Hughes et al. 2003). In other valleys, such as Kato Kambos, there is evidence of a later phase of
666 glaciation. This last phase of glaciation was characterised in this valley and possibly three others
667 (although only the moraine at Kato Kambos is dated) by small cirque glaciers. Exposure ages reveal
668 that this most recent phases of moraine building occurred between. 13 and 10 ka and probably
669 correlate with the Younger Dryas (12.9-11.7 ka). This phase of glaciation was associated with the
670 strongest cirque forms (the four cirque areas are occupied by unit 3 moraines in Fig. 2 and the Phase
671 III glaciers in Fig. 10). In other high valley areas without clearly developed cirques, such as in the
672 upper Neraidhorachi valley at Epano Kambos, where unit 3 moraines are absent (Fig. 2),

673

674 The ^{36}Cl ages are calculated using the excel spreadsheet of Schimmelpfennig et al. (2009) and the
675 authors are aware of continual refinements to the understanding of ^{36}Cl systematics, especially with
676 regard to production rates. Thus, these exposure ages may change in future. Further work is underway
677 to tighten the geochronology by applying U-series techniques to date secondary carbonate cements
678 that are present in the moraines and also to apply optically stimulated luminescence (OSL) dating to
679 glaciofluvial sands. Nevertheless, this paper presents the best available synthesis of the current
680 knowledge of glaciation of Mount Chelmos and as with all research the precise glacial sequence may
681 change as further evidence becomes available. Despite some uncertainties with the stratigraphical
682 succession in some valleys, there is no doubt that: 1) Chelmos was glaciated by an ice cap during the
683 most extensive glaciation; 2) Chelmos was glaciated on multiple occasions, the last being a phase of
684 cirque glaciation, and 3) the Late Pleistocene glaciers reached their maximum earlier than the global
685 LGM which is consistent with many other parts of the Mediterranean mountains.

686

687 In northern Greece, the Late Pleistocene glacial sequence is undated. However, on Mount Tymphi
688 fluvial sediments downstream of the glaciated headwaters of the Voidomatis catchment have been
689 dated and provide some constraint on the upstream glacial chronology (Hughes and Woodward,
690 2008). In this catchment, units comprising glacio-fluvial gravels have been dated to c. 28,200 to
691 24,300 years on the basis of a thermoluminescence date, three electron spin resonance dates and two
692 U-series dates (Lewin *et al.* 1991; Hamlin et al, 2000; Woodward *et al.*, 2008). The sand and silt
693 matrix of this unit is dominated by limestone-rich fines which are derived from glacial erosion in the
694 catchment headwaters (Lewin *et al.*, 1991; Woodward *et al.*, 1992). This geochronology is also

695 supported by findings from the Italian Apennines where the largest glaciers during the Würmian
696 occurred between $31,500 \pm 630$ ^{14}C years BP ($34,770 \pm 638$ cal. years BP) and $22,680 \pm 630$ ^{14}C years
697 BP ($26,239 \pm 789$ cal. years BP) (Giraudi and Frezzotti, 1997). Hughes et al. (2006c, p. 95) argued
698 that “a strong candidate for the last local glacier maximum in the Pindus Mountains is the interval
699 between 30,000 and 25,000 cal. years BP”. This assertion was reiterated by Giraudi (2012). However,
700 the exposure ages of 30-40 ka obtained from Mount Chelmos predate this interval and are consistent
701 with geochronologies showing evidence of a local glacier maximum that predates 30 ka from Iberia
702 (Serrano et al., 2012), the Pyrenees (Delmas et al. 2011) and Morocco (Hughes et al. 2011b). In
703 Turkey there is evidence of a glacier advance pre-dating the global LGM, although glaciers were at
704 their largest at the LGM (Sarıkaya et al., 2014). The same is apparent across the Turkish mountains
705 (e.g. Akçar et al., 2016; Çiner et al., 2016; Sarıkaya et al. 2016) where the LGM advance in MIS 2
706 was the largest advance of the last glacial cycle. Further work is needed to clarify the exact timing of
707 the local glacial maximum on Mount Chelmos and, as with most places listed above from the
708 Mediterranean and, indeed elsewhere around the world (Hughes et al., 2013), the evidence of an early
709 glacier maximum needs testing and refining further. Nevertheless, the ^{36}Cl exposure ages from
710 Chelmos clearly indicate limited ice extent in MIS 2 with a succession of moraines in the highest area
711 of Chelmos yielding the same ages (overlapping within error) which suggests rapid retreat close to the
712 global LGM.

713

714 The most extensive ice cap glaciation was characterised by a single contiguous ice mass that covered
715 an area of nearly 10 km^2 with outlet glaciers extending 1.5 to 3.5 km. The ice mass that submerged
716 the central plateau of Chelmos was constrained by topography and hemmed in the central plateau area
717 by the highest peaks of the massif. This and the fact that ice covered areas of $<100 \text{ km}^2$ means that the
718 ice mass that covered Chelmos is best described as a plateau ice field (cf. Rea *et al.*, 1999; McDougall
719 2001; Evans et al., 2002; Rea and Evans 2003; Vieira 2008). The plateau ice field central dome would
720 have raised the elevation of Mount Chelmos to >2400 m above modern sea level. It is likely that
721 increased areas at high elevation caused by the ice cap surface would have enhanced orographic
722 effects on precipitation. This positive feedback effect would have been important in enhancing glacier
723 development during the most extensive glaciations.

724

725 The average ELA of the largest Middle Pleistocene ice cap and associated outlet glaciers was 1984 m.
726 Two later generations of Late Pleistocene cirque glaciers had average ELAs of 1986 (30-40 ka) and
727 2114 m (Younger Dryas). The largest standard deviations are associated with the Phase I ice cap
728 glacier reconstructions (Table 5) and this reflects large variability in the glacier front altitudes. The
729 smallest standard deviations were associated with the smallest and highest cirque glaciers (Phase III)
730 and this reflects the similar altitudes of the cirque basins on Chelmos. The Phase II glaciers had a
731 similar ELA to the Phase I glaciers when applying an AAR of 0.6. This apparent paradox is because

732 the Phase I ice cap caused an increase in the height of Chelmos since it submerged the landscape. The
733 later Phase II glaciers then occupied incised cirques and valleys inherited from the Phase I glaciation.
734 This meant that even with lower ELAs these glaciers were unable to reach thicknesses sufficient to
735 submerge the mountain centre as a plateau ice field. It is also possible that the ELAs of the plateau ice
736 field and associated outlet glaciers of Phase I were associated with a different AAR to the cirque and
737 valley glaciers of Phase II. For example, Leonard (1984, p. 68) noted that AARs for ice caps may be
738 as high as 0.8. If an AAR of 0.8 is applied to the Phase I plateau ice field, then this corresponds to an
739 ELA of 1854 m a.s.l., more than 100 m lower than the ELA of the Phase II glaciers estimated using an
740 AAR of 0.6. In fact, for Montenegro, an AAR of 0.8 was applied for ice caps on Orjen and Durmitor
741 (Hughes et al. 2010; 2011a) on the basis that this AAR was associated with the AAR with lowest
742 standard deviation of altitude. For Chelmos, this was not the case and knowing which AAR is most
743 appropriate for the different glacier phases is problematic and thus, the ELAs associated with all
744 AARs are shown in Table 5.

745

746 In northern Greece, On Mount Tymphi, the largest Middle Pleistocene glaciers had an average ELA
747 of 1741 m during the Skamnelliian Stage (MIS 12). The Late Pleistocene Tymphian glaciers in this
748 massif had an average ELA of 2174 m a.s.l. Younger Dryas glaciers on Smolikas in northern Greece
749 (they were absent on Tymphi) had an average ELA of 2425 m. The largest Middle Pleistocene
750 glaciation was characterised by ELAs that were 243 m higher in the Peloponnesus than in northern
751 Greece. Conversely, the Late Pleistocene glaciers appear to have been lower in the Peloponnesus than
752 in northern Greece. The Middle Pleistocene comparison is consistent with a warmer climate in the
753 Peloponnesus compared with northern Greece (separated by $> 2^\circ$ of latitude), whereas the Late
754 Pleistocene comparison is the reverse of what one would expect. If the Late Pleistocene comparison is
755 valid, and that Peloponnesus glaciers had lower ELAs than glaciers in northern Greece, then this must
756 be explained by wetter conditions in the former. This is plausible if Late Pleistocene cold stage
757 depressions had a more southerly track than today. A wider comparison, with Montenegro, c. $5-6^\circ$ to
758 the north of the Peloponnesus, illustrates that the Pleistocene glaciers of Greece were significantly
759 higher in all glacier phases (Table 4). For the Late Pleistocene, the Chelmos glaciers were 200-800 m
760 higher than in Montenegro a situation forced not only by lower temperatures in Montenegro, but also
761 higher precipitation, especially on the coastal mountains. The fact that ELAs appear to have been
762 higher in northern Greece than in both Montenegro to the north and the Peloponnesus to the south
763 either has major significance for understanding precipitation sources during the last cold stage *or* is an
764 artefact of tectonic distortion of glacial landform altitudes over the past 30,000 years. As noted earlier,
765 comparisons of ELAs are potentially compromised by different uplift histories (Bathrellos et al. 2016;
766 Giraudi and Giaccio, 2016). Nevertheless, the suggestion that conditions were wetter on Mount
767 Chelmos compared with mountains further north in Greece is consistent with the absence of Late
768 Pleistocene rock glaciers in this massif. Whilst moraines are often substantial in terms of debris

content on Mount Chelmos, they do not resemble the rock glaciers of northern Greece, which are often very clear and abundant (Hughes et al., 2003; Palmentola and Stamatopoulos 2006). Rock glaciers form when debris supply exceeds snow accumulation and thus tend to form under drier conditions than true glaciers (Haeberli, 1985; Belloni et al., 1988).

Tectonic uplift is known to affect the altitudes of Pleistocene glacial landforms in Greece (Bathrellos et al., 2014; Bathrellos et al., 2016) and other Mediterranean mountain areas (Giraudi and Giaccio, 2016). However, whilst tectonics movements are very active in the Northern Peloponnesus area, these are unlikely to affect reconstructed ELAs for the Late Pleistocene. Armijo et al. (1992) and McNeill et al. (2004) argue that Pleistocene and Holocene uplift rates are of the order of 1.5 mm yr^{-1} . This would mean that since the global Last Glacial Maximum (c. 25 ka) Chelmos has been uplifted by less than 40 m. Locally, tectonic uplift may be greater than this, especially around active faults. It is also known that uplift has been greater during the Holocene than during the Pleistocene (Pirazzoli, et al., 2004). The southern coast of the Gulf of Corinth is complicated by numerous local faults (e.g. McNeill and Collier, 2004; Pirazzoli, et al., 2004; Palyvos et al., 2008; Pacchiani & Lyon-Caen, 2010; Ford et al., 2013; including, from North to South, the Pyrgaki-Mamousia, Doumena, Kerpini and the Chelmos Faults and also the East Eliki Fault (all the faults strike East-West). The East Eliki Fault is usually considered as active whilst the southern faults are considered inactive, although there is discussion on the possible activity of Pyrgaki - Mamousia Fault, as there was an earthquake activity in 2004 (Agios Ioannis earthquake of magnitude 4.3) which was correlated with another fault, the Kerinitis Fault (Pacchiani & Lyon-Caen, 2010). The Kerinitis fault has 50°N strike and it is thought to be a fault that connects two segments of the Pyrgaki-Mamousia Fault (Ford et al. 2013). Even with uplift rates estimated from this very active area of 3 mm yr^{-1} for the Holocene (Pirazzoli, et al., 2004) and 1.5 mm yr^{-1} for the Late Pleistocene this would suggest uplift of just 30 m (Holocene) plus 22.5 m for the Late Pleistocene, making a potential total uplift of 52.5 m since 25 ka. Thus, given the small scale of potential uplift relative to the timescales under consideration and the uncertainties associated with providing an accurate value, no correction is warranted to account for uplift for Late Pleistocene glaciers reconstructed ELAs. In fact, since eustatic sea levels were 120 m lower around Greece (Lambeck, 1995), then the reconstructed ELAs (presented as above modern sea level) may be lower than if presented as above palaeo sea level. This would mean that Mount Chelmos was in fact higher above sea level during the Late Pleistocene than is the case today. This phenomenon was also noted for Tymphi in the North Pindus, where Hughes (2004, p. 227) noted that at around the time of the LGM “reconstructed ELAs will underestimate the 'real' ELA by 104 - 112 m”. This calculation was based on an uplift rate for the Epirus region of $0.4\text{-}0.8 \text{ mm yr}^{-1}$ (King and Bailey, 1985). Beyond 300,000 years, uplift outstrips a minimum sea level depression of 120 m and reconstructed ELAs for the Skamnellian Stage (MIS 12) are likely to be overestimated rather than underestimated. Even with the greater uplift rates suggested for the Peloponnesus (1.5 mm yr^{-1}), only beyond c. 80,000 years

806 does uplift outstrip cold stage sea level depressions. Thus, for the oldest and biggest glaciations
807 recorded in Greece, which occurred during the Middle Pleistocene, the altitudinal difference between
808 mountain ranges and sea-level is likely to have been lower than today.

809

810 The implications of the uplift discussion are that glaciers do appear to have been lower than in
811 northern Greece during the Late Pleistocene and that this is not simply an artefact of uplift variations
812 across Greece. This is highly significant because it implies that conditions in the Peloponnesus were
813 more conducive for glaciation than in the northern Pindus. This is unlikely to be explained by
814 temperatures but instead suggests that conditions were wetter in the Peloponnesus than in northern
815 Greece during the glacier advances associated with the Late Pleistocene glacier maximum and also
816 the Younger Dryas. The fact that further north, in Montenegro, glacier ELAs are substantially lower
817 than in both northern and southern Greece, suggests that here the conditions were not only colder but
818 wetter too. This implies that moisture bearing atmospheric systems that delivered winter snow to
819 west-central Balkans and southernmost Greece were not the same. Winter precipitation in the west-
820 central Balkans is likely to have been caused by cyclogenesis in the northern Adriatic (Hughes et al.
821 2010) whereas in the Peloponnesus moisture-bearing atmospheric systems must have had a western or
822 southern source. Whilst this may seem counter-intuitive with regard to air temperatures it is moisture
823 which is most important; and in a Pleistocene cold stage winter even southerly air masses would be
824 sufficiently cold to deliver snow. This is consistent with recent findings from the Alps that
825 depressions had a southerly track across the southern Mediterranean basin which then veered
826 northwards approaching the Alps from the south (Luetscher et al. 2015). The trajectory of the jet
827 stream over southern Europe indicated by Luetscher et al. (2015, their Fig. 1) is consistent with both
828 cyclogenesis in the western Mediterranean (caused by a southerly jet stream source in this area) and
829 also cyclogenesis in the northern Adriatic in the lee-side of the eastern Alps (caused by a northerly jet
830 stream source in this area). However, further work is needed to constrain the timings of glaciations in
831 both northern Greece and the west-central Balkans since in these areas Late Pleistocene moraines are
832 either not dated or not precisely dated since U-series techniques provide only minimum ages (Hughes
833 et al. 2010; 2011a). The success of ^{36}Cl in dating late Pleistocene moraine surfaces in the
834 Peloponnesus as demonstrated in this paper will hopefully provide the incentive to now date other
835 moraines across Greece and the west-central Balkans where limestone terrains dominate.

836

837 There is evidence that glaciers were present on other Peloponnesus mountains during the Pleistocene.
838 Mastronuzzi et al. (1994) reported moraines in the cirques and valleys of the Taygetos (2404 m a.s.l.).
839 In a large north-facing cirque (Actoctionios) multiple moraines are present down to 1500 m a.s.l. In the
840 eastern cirques moraine crests are higher, situated between 1840 and 1880 m a.s.l. However, glacial
841 landforms are also present below these moraines with roche moutonnées present at 1740 m a.s.l. and it
842 is possible that moraines are present in forested areas below the treeline. Pope (2013) noted that these

glaciated catchments feed extensive fan formations in the Sparta area. Luminescence dating suggests that proximal fan sediments were deposited between 250 ka to 130 ka (MIS 8 to 6) (Pope and Wilkinson, 2006). However, the moraine and proximal fan soils differ significantly in terms of magnetic and iron properties, which are proxies of weathering (Pope and Millington 2000), and this may reflect a large age difference. The relationship between the Taygetos glaciations and the Sparta fans is the subject of ongoing research by the authors. Elsewhere in the Peloponnesus, little is known about the glacial history of the mountains. It is likely that glaciers comparable with those on Chelmos and the Taygetos are also present on Erimanthos (2124 m a.s.l.) to the west and Kyllini (2374 m a.s.l.) to the east.

Conclusions

There are at least three glacial phases recorded in the glacial geomorphology of Mount Chelmos. The oldest and largest is undated and is correlated with the large Middle Pleistocene glaciations recorded elsewhere in Greece and the wider Balkans. The moraines in the highest cirques have been dated, using ^{36}Cl exposure dating, and provide important constraints on the timings of Late Pleistocene glacier advance and retreat in Greece. The largest glaciers of the Late Pleistocene formed before 30 ka with less extensive recessional moraines dating close to the global LGM. Glaciers were therefore likely to have been present for over 10 ka from MIS 3 through to MIS 2. Cirque moraines post-date the earlier moraines and are present in all the cirques of Chelmos. The ^{36}Cl exposure ages of 12.6 ± 0.9 and 10.2 ± 0.7 ka are consistent with a Younger Dryas age for the last glaciers of Mount Chelmos. The Late Pleistocene moraine chronology from Mount Chelmos provides preliminary insight into glacier history of Greece during the last cold stage. The new ^{36}Cl exposure ages not only confirm earlier suggestions of an early glacier maximum, but also indicate the presence of glaciers in this area through the global LGM and also during the Younger Dryas. The absence of Late Pleistocene rock glaciers on Mount Chelmos may indicate wetter conditions in this area compared with further north in Greece, where rock glaciers are abundant.

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1284 Figures

1285 Figure 1. Location of Mount Chelmos in Greece showing the major tectonic features of the
1286 area.

1287
1288 Figure 2. Glacial geomorphological map of Mount Chelmos showing sample locations and
1289 exposure ages in red font.

1290
1291 Figure 3. Glacial boulders of Unit 1 in the Spanolakos valley. These represent the oldest
1292 glacial deposits in this valley. The boulders rest on cemented sands and gravels.

1293
1294 Figure 4. Clear arcuate end moraine in the Spanolakos valley (Unit 2).

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1296 Figure 5. Boulders sampled for ^{36}Cl exposure dating on a lateral moraine crest in the Kato
1297 Kambos valley. Foreground boulder is CH8 (with hammer) and the boulder near the person
1298 in the background is CH7. These boulders yielded exposure ages of 30.4 ± 2.2 ka and $39.6 \pm$
1299 3.0 , respectively.

1300
1301 Figure 6. The cirque at Kato Kambos with bounding moraine ridge highlighted by sunlight
1302 (Unit 3).

1303
1304 Figure 7. Boulder sampled for exposure dating the Kato Kambos cirque (Unit 3).

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1306 Figure 8. The lowest moraines in the Chaliki valley.

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1308 Figure 9. Cirque moraines southeast of spot height 2318 m (peak on the left side of
1309 photograph).

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1312 Figure 10. The three glacier phases on Mount Chelmos showing their relative sizes.

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1315 Tables

1316

1317 Table 1. Sample details for ³⁶Cl analysis.

1318

Sample	Altitude (m)	Latitude (°N)	Longitude (°E)	scaling factor for nucleonic production ¹	scaling factor for muonic production ¹	Thickness	Density ²	Shielding ³
CH2	2250	37.9436	22.1969	5.1794	2.4875	1	2.5	0.995
CH4	2221	37.9757	22.1973	5.0839	2.4598	3	2.5	0.960
CH5	2209	37.9765	22.1973	5.0434	2.4481	2	2.5	0.960
CH7	2120	37.9674	22.1900	4.7493	2.3620	2	2.5	0.960
CH8	2121	37.9674	22.1903	4.7525	2.3629	5	2.5	0.950
CH10	2140	37.9655	22.1927	4.8137	2.3810	2	2.5	1.000
CH11	2139	37.9656	22.1927	4.8105	2.3801	5	2.7	0.980

1319

1320 ¹Calculated following scaling factors and equations in Stone (2000).

1321 ²Measured using water displacement.

1322 ³Calculated using Cronus web calculator, available at

1323 http://hess.ess.washington.edu/math/general/skyline_input.php

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1325

Table 2. Geochemical data for the ³⁶Cl samples.

Sample	Major elements											Trace elements					
	CaO [wt-%]	K ₂ O [wt-%]	TiO ₂ [wt-%]	Fe ₂ O ₃ [wt-%]	SiO ₂ [wt-%]	Na ₂ O [wt-%]	MgO [wt-%]	Al ₂ O ₃ [wt-%]	MnO [wt-%]	P ₂ O ₅ [wt-%]	CO ₂ [wt-%]	Li [ppm]	B [ppm]	Sm [ppm]	Gd [ppm]	Th [ppm]	U [ppm]
CH2	47.870	0.488	0.137	1.042	8.805	0.112	0.728	2.601	0.068	0.108	37.900	0.173	35.916	0.687	0.953	1.142	2.427
CH4	57.513	0.049	0.020	0.026	1.956	0.023	0.264	0.284	0.052	0.157	39.300	0.040	89.722	0.761	0.989	5.193	3.456
CH5	47.048	0.499	0.160	1.090	9.401	0.107	0.079	2.824	0.059	0.070	37.900	0.099	74.670	1.445	1.579	2.573	5.085
CH7	49.662	0.423	0.130	1.058	7.560	0.117	0.907	2.293	0.067	0.083	37.600	0.111	35.621	0.771	0.981	1.562	6.158
CH8	52.708	0.255	0.080	0.603	4.308	0.086	0.668	1.486	0.046	0.058	39.600	0.069	83.543	1.500	1.744	2.028	4.463
CH10	54.174	0.165	0.062	0.440	5.831	0.076	0.540	1.090	0.077	0.059	37.400	0.064	83.734	1.403	1.887	1.919	2.972
CH11	39.533	0.086	0.035	0.416	46.128	0.022	0.233	0.579	0.079	0.187	12.600	0.150	9.838	2.924	3.111	1.136	1.097

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1327 | Table 3. AMS measurements of ³⁶Cl and associated exposure ages calculated using the excel
1328 calculator of Schimmelpfennig et al. (2009). Order in descending order of calculated
1329 exposure age.

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Sam ple nam e	Sam ple weig ht (g)	Add ed spik e (mg)	³⁶ Cl/ Cl (e- 15)	Erro r (e- 15)	³⁵ Cl/ ³⁷ Cl	Erro r	Cl (pp m)	Erro r (%)	³⁶ Cl atom / sam ple (g)	Erro r %	³⁶ Cl expo sure age (ka)	Erro r (ka)
CH1 1	33.7 353	0.92 75	863	20	8.23 9	0.09 3	20.4 00	2.1	7135 24	2.49	10.1 56	0.72 0
CH1 0	30.5 065	1.01 08	1580	25	13.7 84	0.00 8	11.3 81	1.6	1224 632	1.61	12.6 13	0.87 3
CH4	32.0 071	1.02 10	2171	71	8.62 5	0.01 0	22.5 62	3.3	2048 733	3.28	21.1 52	1.62 8
CH5	31.8 594	1.01 97	2450	55	16.1 19	0.02 3	8.77 4	2.2	1746 687	2.24	21.5 86	1.55 3
CH2	30.4 977	0.99 38	2748	44	12.8 08	0.11 4	11.8 00	1.9	2152 481	1.65	22.8 50	1.60 6
CH7	30.5 769	1.02 34	4307	119	14.2 73	0.05 2	11.7 61	2.8	3345 379	2.76	39.8 50	3.01 3
CH8	31.7 567	1.02 50	3308	75	12.3 33	0.03 3	13.0 70	2.3	2616 728	2.28	30.4 17	2.24 7

1332 Table 4. ³⁶Cl exposure ages relative to the overall glacial stratigraphy and possible
1333 correlations with wider climatic events and stratigraphical units. No geochronology is
1334 available for the lowermost glacial deposits (Unit 1) and a Middle Plietsocene age is
1335 hypothesised based on correlations with dated moraines in northern Greece (Hughes et al.
1336 2006).

1337

Stratigraphical Unit	³⁶ Cl Exposure Ages		Correlations (tentative)
4	12.6 ± 0.9 [Kato Kambos] 10.2 ± 0.7 [Kato Kambos]	End moraine	Younger Dryas Glacier stabilisation/advance
3	21.2 ± 1.6 [Epano Kambos] 21.6 ± 1.6 [Epano Kambos] 22.9 ± 1.6 [Epano Kambos]	Recessional moraines	LGM Glacier retreat
2	39.9 ± 3.0 [Kato Kambos] 30.4 ± 2.2 [Kato Kambos]	End moraine	MIS 3 Glacier stabilisation/advance
1	Undated	Moraines and till down-valley of unit 2	Middle Pleistocene

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1340 Table 5. Potential Equilibrium Line Altitudes (ELAs) associated with a range of
 1341 Accumulation Area Ratios (AARs) for the three glaciers phases on Mount Chelmos. The
 1342 most appropriate AAR may depend on the geometry and hypsometry of the former glaciers.
 1343 For example, the ELAs of the ice cap and outlet glaciers associated with Phase I may not
 1344 have been associated with the same AAR as the cirque and valley glaciers of Phases II and
 1345 III. See main text for discussion.

PHASE I	Spanolakos	Strogilolaka	Kato Kambos	Chaliki	Neraidhorachi	Xerokambos	TOTAL AREA	
Area (km ²)	1.014	0.954	0.954	1.489	3.156	2.171	9.738	
AAR	Altitudes (m)						Mean	St Dev
0.4	2170	2240	2155	1900	2200	2120	2131	120.2
0.5	2110	2195	2145	1840	1980	2065	2056	128.6
0.6	2060	2160	2130	1765	1770	2020	1984	175.0
0.7	2010	2130	2120	1700	1610	1960	1922	218.3
0.8	1940	2090	2100	1625	1480	1890	1854	251.9
PHASE II	Spanolakos		Kato Kambos	Chaliki	Neraidhorachi	Xerokambos	TOTAL AREA	
Area (km ²)	0.365		0.676	0.802	1.623	0.979	4.445	
AAR	Altitudes (m)						Mean	St Dev
0.4	2060		2200	2025	2025	2020	2066	76.6
0.5	2050		2175	1925	2000	1990	2028	93.4
0.6	2040		2160	1800	1970	1960	1986	131.1
0.7	2030		2150	1695	1925	1940	1948	167.4
0.8	2015		2140	1645	1900	1910	1922	182.7
PHASE III	Spanolakos		Kato Kambos	Chaliki		Xerokambos	TOTAL AREA	
Area (km ²)	0.134		0.417	0.131		0.050	0.866	
AAR	Altitudes (m)						Mean	St Dev
0.4	2070		2186	2150		2120	2135	50.3
0.5	2060		2178	2145		2065	2125	53.3
0.6	2045		2172	2125		2025	2114	55.3
0.7	2035		2164	2110		1960	2101	59.4
0.8	2020		2155	2090		1890	2085	62.0

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39.9 ± 3.0 ka

30.4 ± 2.3 ka

12.6 ± 0.9 ka

10.2 ± 0.7 ka

12.6 ± 0.9 ka

Glacier reconstruction