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1 **LGM glaciers in the SE Mediterranean? First evidence from glacial landforms and ³⁶Cl dating**
2 **on Mount Lebanon**

3

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17 **Highlights:**

- 18 - First evidence of LGM glaciers in the southeastern Mediterranean (Mount Lebanon;
19 34.25°N) are provided
- 20 - The geometry of three glaciers is reconstructed and constrains the LGM Equilibrium
21 Line Altitude at about 2400 m a.s.l.
- 22 - Glacial and climate modeling indicates a drier LGM climate than today

23

24

25 **Keywords:** Mount Lebanon, Glacial morphology, ³⁶Cl-dating, Levant paleoclimate

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27

28 **Abstract:**

29 The hydrological conditions during the Last Glacial Maximum in the Eastern
30 Mediterranean have long been debated. In particular, very little is known about the
31 development of high-altitude glaciers in this region. In the present study, morphological and
32 sedimentological evidence, such as U-shape valleys, glacial cirques, and morainic ridges, are

33 used to identify past glaciers on the western slopes of Mount Lebanon, which is by far the
34 highest topography along the eastern Mediterranean coast (3088 m a.s.l.). The geometry of
35 three paleo-glaciers is reconstructed: one 8-km-long glacial tongue stretching from the
36 highest peak down to an elevation of about 2000 m a.s.l., and two small glacial cirques located
37 at about 2300 m a.s.l. The age of maximum glacier advance is constrained from a 10-m-deep
38 ^{36}Cl depth-profile within the diamicton of one terminal moraine, and yields an early LGM
39 timing (between 25.4 ± 3.1 ka and 31.2 ± 4.5 ka), which agrees with the chronology of the last
40 maximum glacier extension in the Eastern Mediterranean. The three glaciers together form a
41 consistent group, compatible with an Equilibrium Line Altitude (ELA) at 2400 m estimated
42 using the Accumulation Area Ratio method. A degree-day model is further used to infer the
43 temperature and precipitation conditions required to depress the LGM ELA to 2400 m. Our
44 results suggest that the observed glacier extent is compatible with early LGM paleo-
45 temperature records of the area only under the condition of decreased precipitation relative
46 to modern times. These findings thus support an early LGM climate of the Levant region that
47 was drier than today.

48

49 **1. Introduction**

50 The Last Glacial Maximum (LGM) corresponds to the last time interval in Earth's history,
51 when the globally-integrated volume of ice sheets reached its maximum under the influence
52 of external forcing (Mix et al., 2001). The timing of the LGM varies depending on the type of
53 used geological record (e.g. between 26 and 19 ka from globally-averaged sea-level records
54 (Clark et al., 2009) versus between 28 and 23 ka from atmospheric dust flux recorded in polar
55 ice cores (Hughes and Gibbard, 2015)). In any case, the mode and rate at which external
56 forcing mechanisms are transmitted to Earth's surface and control local climates are neither
57 homogeneous nor fully understood. Reconstructing the distribution of LGM climate
58 parameters in space and time with sufficient resolution is thus essential to solve for the spatial
59 variability of the LGM climate. Moreover, these reconstructions allow for investigation of the
60 feedbacks involved in the transition from glacial to interglacial periods in areas located far
61 away from polar regions, thus improving our understanding of modern climate dynamics, as
62 well as our ability to forecast climate evolution. Local paleoclimate reconstructions are
63 fundamental to disentangle how global forcing is modulated at regional/local scales, how local
64 climates interact with each other, and thus permit to improve large-scale climatic models.

65 The Mediterranean region is highly sensitive to global climate change. Paleoclimatic
66 records indicate that the LGM in the Mediterranean was marked by a northwestward gradient
67 of both the atmospheric temperature anomaly (relative to the present), reaching a maximum
68 of -17°C in the Alps (Wu et al., 2007) (pollen-data-based reconstructions), and the sea surface
69 temperature anomalies (Mikolajewicz, 2011). Superimposed on these regional trends are
70 glacial refugia, where the LGM climate was locally buffered, as indicated by the persistence of
71 endemic species in climatic niches over geological times (e.g. Médail and Diadema, 2009; Petit
72 et al., 2003), making the Mediterranean a hotspot of biodiversity.

73 Here, we study a new LGM climatic record from Mount Lebanon (coastal Levant,
74 Eastern Mediterranean), a mountainous environment system where local responses to
75 climate forcing are expected. The Levant region (Figure 1b) is located in a transitional zone
76 between the temperate Mediterranean climate to the North and the sub-tropical high-
77 pressure zone to the South (Hajar et al., 2010; Develle et al. 2010). Its modern climate is
78 typically Mediterranean, on which topographic and continental effects are superimposed
79 locally due to the presence of high mountain ranges (Mt Lebanon 3087 m a.s.l., Figures 1b and
80 2a). Paleotemperature reconstructions based on several proxies suggest a mean annual LGM
81 temperature difference (relative to present) ranging between -6 to -14°C indicating a broad
82 agreement for a regional glacial cooling (Affek et al., 2008 ; Almogi-Labin et al., 2009; Emeis
83 et al., 2003; Essallami et al., 2007; McGarry et al., 2004; Wu et al., 2007). The main debate
84 about the glacial climate of the Levant region deals with the question whether it was dry or
85 wet (Develle et al. 2010; Miebach et al., 2019). Although paleohydrological records show
86 pronounced changes associated with the last glacial stage across the whole region, there is a
87 marked disparity in the spatial response between the central and southern Levant. In the
88 central Levant, records from the Yammouneh paleo-lake in Lebanon indicate that the LGM
89 was a period of reduced water balance of the lake (Develle et al., 2010; Gasse et al., 2015). In
90 the southern Levant on the other hand, some 150 km further south in Israel (Figure 1b), soil
91 sequence stratigraphy was used by Gvirtzman and Wieder (2001) to infer wet LGM conditions.
92 LGM lake high-stands in the Lisan Dead Sea Basin have also long been attributed to increased
93 precipitation (Stein et al., 2010; Torfstein et al., 2013a,b), although this latter interpretation
94 has been recently questioned (Miebach et al., 2019; Stockhecke et al., 2016).

95 The present study focuses on glacial deposits, a type of continental record that can be
96 more robustly interpreted in terms of temperature and precipitation, as the

97 advances/retreats of glaciers respond quickly to changes in the local Equilibrium Line Altitude
98 (=ELA) (Porter, 2000) and hence to the vertical structure of the atmosphere (Braithwaite,
99 2008). The ELA is the elevation where the annual mass balance of snow/ice is zero, and above
100 which ice forms by accumulation of snow over years (Porter, 2000). The size of glaciers and
101 their downward extent are therefore strongly regulated by the local temperature and
102 precipitation (Blard et al., 2009; Hughes et al., 2010; Jomelli et al., 2009; Oerlemans, 2005;
103 Protin et al., 2019; Rabatel et al., 2008), although secondary factors such as slope orientation,
104 wind, avalanches, can also contribute significantly to the distribution of glaciers (Dahl and
105 Nesje, 1992; Farinotti et al., 2020; Hewitt et al., 2011). Around the Mediterranean, large LGM
106 glaciers concentrate along an ENE-WSW-trending line running from the Carpathian (Romania)
107 to the Cantabrian (Spain) Mountains, and culminating in the Alpine ice sheet (LGM ELA
108 between 1000 and 1500m) (Figure 1a). Based on erosional and depositional features, evidence
109 of small LGM glaciers were also found locally in other unfavorable (more southern) circum-
110 Mediterranean areas, where topographic highs exceeded the LGM ELA (i.e. in Morocco, Spain,
111 Italy, Balkans, Greece, and Turkey) (Kuhlemann et al., 2008, Hughes and Woodward, 2017)
112 (Figure 1a). All so far known LGM Mediterranean glaciers are located NW of a line connecting
113 eastern Turkey to central Morocco (Allard et al., 2021; Allen et al., 2008; Ehlers and Gibbard,
114 2004; Hughes and Woodward, 2017; Kuhlemann et al., 2008) (Figure 1a). The only known LGM
115 glaciers south of latitude 37°N are located on the Atlantic seaboard, on the up-to-4000-m
116 topography of the Atlas Mountains (Morocco; LGM ELA between 2400 and 3400m) (Figure
117 1a). It should be acknowledged however that the preservation of morainic deposits on Mount
118 Lebanon and Mount Hermon (Lebanon) was reported by Messerli (1966). If confirmed, the
119 existence of these glacial records at this latitude (34°N) of the Levant region would extend the
120 southeastern limit of former Mediterranean glaciers by several hundreds of kilometers (Figure
121 1a).

122 The basic strategy when reconstructing paleoclimate parameters from glacial deposits
123 is to identify terminal moraines (i.e. the sedimentary formations left in front of a glacier when
124 the latter starts retreating after a phase of equilibrium; Refsnider et al., 2007), to date them,
125 and then to calculate the ELA required to allow the glacier advancing down to that terminal
126 moraine (Blard et al., 2009; Protin et al., 2019). The ELA is essentially a bivariant system, which
127 in most cases, highly depends on the temperature, and in a lesser extent on the rate of
128 precipitation (e.g. Allen et al., 2008). Here, we show through geomorphic mapping and ³⁶Cl

129 exposure dating, that Mount Lebanon hosts glacial features and sediments that testify to the
130 development of glaciers during the LGM in agreement with Messerli (1966), and we discuss
131 how these results translate into the LGM climatic conditions of the Levant region.

132

133 **2. Geomorphic and climatic setting of Mount Lebanon and the Cedars cirque**

134 Mountain ranges in Lebanon (latitude $\approx 34^\circ\text{N}$) formed as a result of compressive
135 deformation across the Lebanese restraining bend of the Dead Sea Transform plate boundary
136 (Elias et al., 2007) (Figure 1b). Sharply bounded to the E by the Yammouneh fault, which is
137 part of the Dead Sea plate boundary, Mount Lebanon is a 150-km-long NNE-SSW-trending
138 mountain range. Both its width and elevation increase northeastward (Figure 2a). The modern
139 climate over Mount Lebanon results from the superimposition of the orographic effect
140 (elevation up to 3000 m) on the Mediterranean influence (hot-dry summers and cool-wet
141 winters). The monthly average temperature in Bcharre (the meteorological station closest to
142 the studied sites; see Figure 2a), 1430 m a.s.l., ranges from 0°C in January and February to
143 17°C in July and August. Annual precipitation is 924mm, increasing up to 1400 mm in Mount
144 Lebanon (Figure 2a) (Cheddadi and Khater (2016) after Plassard (1981)), 85% of which occurs
145 between November and March (Abi-Saleh and Safi, 1988). The impact of the orographic effect
146 on the rates of precipitation is evidenced by the sharp decrease in precipitation between
147 Mount Lebanon and the semi-arid to arid Bekaa Valley (Figure 2a).

148 Our study area is located on Jabal El Makmal, the highest mountain of the Eastern
149 Mediterranean – Levant coast culminating at around 3080 m a.s.l. in Qornet Es Sawda. The
150 mountain top extends southwards as a ~ 70 km plateau – Mnaïtra Plateau (Figure 2a).
151 Structurally, this plateau corresponds to the flat limb of the west-dipping Mt-Lebanon
152 monocline. The plateau bounds the Yammouneh fault and plate boundary along a 500-700-
153 m-high escarpment separating Mt Lebanon from the Bekaa Valley (MLBV escarpment)
154 (Daëron et al., 2007) (Figures 2a–c). The surface of the plateau is carved into the 500-700-m-
155 thick Sannine limestone formation of the early Upper Cretaceous (Walley, 1997), and is
156 exposed to dominant winds blowing from the W (Maheras, 1988; Messerli, 1966). Its surface
157 displays a poorly-channelized morphology marked by prominent evidence of karstic erosion
158 (numerous dolines in particular). When exposed at the western rim of the Mnaïtra plateau,
159 the hard Sannine formation forms a prominent topographic step that follows a SSW-NNE
160 trend parallel to both the Mediterranean shoreline and the axis of the monocline (Figure 2a).

161 This escarpment is locally dissected by the headward retreat of Mediterranean tributaries
162 (Figure 2a). Near the highest elevations of Mt Lebanon, the rim of the plateau deviates from
163 its SSW-NNE general trend, as it bends in the E-W direction along the upper course of the
164 Qadisha River, to form the walls of the Cedars cirque (Figures 2a–c).

165 The Cedars cirque is a major horseshoe-shaped incision (in map view), open to the
166 west, extending for 10-12 km in E-W direction over the entire width of the Jabal el Makmal
167 relief. It is 6-7 km wide with an area of about 28 km² (Figures 2a–c). The floor of the cirque
168 has an average elevation of 1900m, and is dissected by a 300-400-m-deep, E-W-trending
169 canyon of the Qadisha river. It is separated from the plateau by a laterally continuous 600-
170 900-m-high escarpment made of the resistant limestone of the Sannine formation (Figures 2b,
171 3b–d). The steepest topographic profile of this escarpment is generally concave upwards, with
172 the slope increasing from 5° at the base to 30° at the top (Figures 2b and 4a). The sloping
173 surface of the escarpment is relatively smooth and fairly continuous, covered by numerous
174 colluvial fans (Figures 2c, 3c-d). According to the present topography, there is only one point
175 of hydrological connection between the plateau and the cirque. It corresponds to the outlet
176 of a 4-5-km-long, flat bottom U-shaped valley (hereafter referred to as the Sawda valley;
177 Figures 3a and 3e) floored by sorted clasts typical of periglacial conditions (i.e. upfreezing,
178 Anderson (1988)). The Sawda valley occupies a special position within the Mnaitra plateau
179 since it is the only thalweg which is not oriented along the NNW-SSE regional steepest
180 gradient. Its upper reach is located on the SE side of the Qornet Es Sawda (Figure 2c), making
181 the Sawda valley the only leeward portion of the plateau and hence favoring accumulation of
182 ice there (Messerli, 1966). The valley is oriented parallel to the MLBV escarpment (Figure 2c)
183 and is characterized by a very gentle topographic gradient (less than 1°) at elevation between
184 about 2950 and 2800m a.s.l. It terminates at the NE edge of the Cedars cirque in the form of
185 a 100-150-m-deep U-shaped notch suspended above the cirque floor (Figures 2c, and blue
186 arrow in Figure 3b)

187 Field observations and the analysis of 1:20000 topographical maps and 2.5-m-
188 resolution SPOT images allowed a distinctive top-down series of landforms, characteristic of
189 glacier formations, to be mapped at three distinct locations within the Cedars cirque
190 (Manhale, Qandîl, and Terkmén sites in Figure 2). Each series hosts one, or a group of few, 10-
191 m-high 100-m-wide curvilinear morainic ridges (Figures 3b, 3d, 3e, and 3f) that separates a
192 depressed clay-rich level terrain uphill (Figure 3b, 3d, and 3f) from a field of hummocks

193 downhill (Figure 3c). However, despite these similarities, there is a fundamental difference
194 between Manhale and the two other sites. At Terkmen and Qandîl, the typical landform
195 succession rests on the steep cirque wall (Figures 2b and 2c), such that the topographic cross-
196 section features a prominent abrasional hollow isolated by a downstream ridge (Figure 4b).
197 This clay-rich depression is almost topographically closed, and possibly reveals an ancient pro-
198 glacial lake (Figures 2c and 3b). At Manhale, however, the landform succession covers the
199 gently sloping floor of the Cedars cirque, such that the topographic signature is more subdued
200 and the clay-rich level terrain has a broader extension in the upstream direction (Figures 2c,
201 3d, 3f, and 4b). In addition, the Manhale succession lays immediately below the U-shaped
202 notch outlet of the Sawda valley (Figures 2c), which suggests that the Manhale moraine might
203 have been emplaced by a former glacier originating from the Sawda valley and cascading on
204 the northeastern wall of the Cedars cirque.

205 At Manhale and Terkmen (but not at Qandîl), large pieces of limestone conglomerates,
206 composed of moderately to well-rounded 1-50-cm pebbles and cemented by secondary
207 calcite, are exposed at the surface of the ridges (Figure 3h). These characteristics are
208 commonly observed at the surface of limestone moraines (Hughes et al., 2010). The core of
209 the westernmost Manhale ridge, exposed in a large man-made excavation, is a matrix-
210 supported, unsorted, and unstratified diamicton, composed of rounded to sub-rounded
211 cobbles and pebbles of limestones floating within a white silty matrix (Figure 5). Boulders are
212 only occasionally found and never exceed 1m. The upper cemented horizon is thin, and does
213 not propagate deeper than a few tens of centimeters below the surface (Figure 5). Although
214 the core of the Terkmen ridges is not excavated, we assume, based on the similar surface
215 morphology and composition, that it is composed of the same diamicton as that in Manhale.
216 On the other hand, the absence of the cemented conglomerate at the surface of the Qandîl
217 ridge (see different color code in Figure 2c) suggests the latter might represent a bedrock
218 abrasional ridge without deposition.

219 The clear evidence of abrasional landforms (U-shaped Sawda valley, and small cirques
220 at Terkmen and Qandîl) at an elevation compatible with known LGM glaciers in climatically
221 similar regions nearby (e.g. Sarikaya et al., 2009) (Figure 1a), and their association with
222 diamicton sediments typical of glacial tills, strongly suggest that the landform successions
223 observed within and above the Cedars cirque are of glacial origin. Several possible glacial
224 origins have been proposed for hummocks, such as the melting of detached patches of ice

225 during glacier retreat (e.g. Bennett and Boulton, 1993; Bennett, 1994), or the subglacial
226 deformation of older till deposits under the overburden pressure of stagnant ice (Boone and
227 Eyes, 2001; Hodgson, 1982). However, in the present case, where karst is likely to play a major
228 role, it is also possible that the hummocky topography formed as a result of suffusion
229 processes, during which till sediments are evacuated through underlying karstic pipes (Çiner
230 et al., 2019; Ford and Williams, 2013).

231 That the Mount Lebanon hosts evidence of past glaciation is consistent with earlier
232 work of Messerli (1966) in this area. Given their orientation relative to the surrounding
233 topography (Figure 2c), the Manhale, Terkmen, and Qandil ridges most likely represent frontal
234 moraines, which thus record a maximum extent of the former glaciers (Refsnider et al., 2007).
235 In the next section, we focus on the Manhale morainic ridges, for which we attempt to derive
236 an age of emplacement through ^{36}Cl Cosmic Ray Exposure dating.

237

238 3. ^{36}Cl Cosmic Ray Exposure Dating of the Manhale ridge

239 The chronology of emplacement of the Manhale ridge was established through ^{36}Cl
240 Cosmic Ray Exposure (CRE) dating of the diamicton deposit. ^{36}Cl is a cosmogenic nuclide
241 produced by the interactions between cosmic ray particles and specific target elements, the
242 most abundant ones in limestone being Ca and Cl. Like any other cosmogenic nuclide, the
243 variability of the rate of ^{36}Cl production at the Earth's surface essentially depends on the
244 latitude, altitude, and topographic shielding of the sampled site, and can be determined from
245 scaling schemes (e.g. Stone, 2000). ^{36}Cl is also a radioactive nuclide, with a decay constant of
246 $2.30 \times 10^{-6} \text{ yr}^{-1}$.

247 For a given site, the concentration of ^{36}Cl in a surface sample is a function of the local
248 production rate (P), the age of surface exposure (t), the rate of denudation (ϵ), and the ^{36}Cl
249 inheritance (C_0) (i.e. the ^{36}Cl inventory contained in the sample at the time of emplacement of
250 the sampled landform). Deriving robust constraints on the age of exposure therefore requires
251 to simultaneously solve for the three latter unknowns. The best way to achieve this is to collect
252 several samples along a depth profile (Braucher et al., 2009; Hidy et al., 2010). This is because
253 secondary neutrons and muons, which are the cosmic ray particles responsible for the
254 cosmogenic production, are attenuated with depth following a specific pattern as a function
255 of denudation and inheritance (Schimmelpfenning et al., 2009). Any collection of

256 (concentration, depth) data thus has the potential of constraining the most likely (t , ϵ , C_0)
257 triplet associated with the sampled landform (Braucher et al., 2009; Hidy et al., 2010).

258 For this purpose, a depth profile was sampled along the vertical subsurface wall of a
259 newly excavated construction site in the Manhale ridge (Figure 5). Eighteen samples, each
260 made of 10-20-cm-large individual pebble, were collected at depths between 0 and 975 cm
261 beneath the ridge axis in order to minimize surface denudation. The major source of
262 uncertainty in modeling cosmogenic depth profiles resides in the estimation of the density of
263 the material, across which the cosmic ray particles are attenuated (Braucher et al., 2009). For
264 this, we first measured the density of the clay matrix by weighing a 1-dm³-volume of matrix in
265 the field. The measured matrix density is 1.2 g cm⁻³, approximately half the 2.7 g cm⁻³ density
266 of the pebbles. We have then estimated the proportion of pebbles within the diamicton to
267 about 30% according to field photographs of the outcrop, yielding a bulk density of
268 approximately 1.65 g cm⁻³.

269 Pebble samples were crushed, and the 250-500- μ m fraction was leached, then
270 dissolved following the procedure of Stone (1996), modified by Schlagenhauf et al. (2011).
271 After complete dissolution, the samples were spiked with a known amount of isotopically
272 enriched stable chloride carrier to allow for simultaneous Cl and ³⁶Cl determination by isotope
273 dilution (Schlagenhauf et al. (2011) and references therein). Cl was then precipitated as AgCl.
274 Both the ³⁶Cl and Cl concentrations were measured at the national French AMS facility ASTER
275 (CEREGE, Aix-en-Provence), and normalized to a ³⁶Cl standard prepared by K. Nishiizumi
276 (KNSTD1600, with a given ³⁶Cl/³⁵Cl value of $(2.11 \pm 0.06) \times 10^{-12}$; Sharma et al., 1990; Fifield et al.
277 1990). The analytical uncertainties included counting statistics, machine stability and blank
278 correction. Calcium concentration was measured in aliquots by using ICP-AES analyses.

279 The obtained ³⁶Cl concentrations vary from 6.267×10^4 to 1.548×10^6 atoms per gram of
280 rock (Figure 5 and Table 1) with a clear exponential decrease of ³⁶Cl concentrations with depth
281 (Figure 5), suggesting that deposition occurred over a time interval that is short relative to the
282 time of exposure, and also that there has been insignificant vertical mixing of pebbles within
283 the diamicton. In other words, the deposit was built over a discrete time, and it has remained
284 relatively stable since then. The depth profile is modeled using a recently developed ³⁶Cl-
285 exposure-calculator (Schimmelpfennig et al., 2019), which makes use of Markov chain Monte
286 Carlo sampler of Goodman and Weare (2010) to construct posterior distributions of the model
287 parameters (t , ϵ , C_0). A value of 48.8 ± 4.8 atoms g⁻¹ a⁻¹ (Stone et al., 1996) was used as the

288 spallation production rate at sea level and high latitude, which was scaled to the sampled site
289 using the Lal (1991), modified by Stone (2000), scaling scheme. Muonic production
290 calculations are performed using the equations of Heisinger et al. (2002a, 2002b) adopting the
291 approach of Balco et al. (2017) to calculate muon production at a particular altitude and
292 subsurface depth. The calculation includes a mean chemical composition for all the samples
293 (Supplementary Table 1) derived from the analysis of 5 samples over the whole profile. The
294 production rate along this profile is reasonably well constrained owing to the low Cl content
295 of the sampled pebbles (Table 1). For the field-estimated density of 1.65 g cm^{-3} , the depth-
296 profile modeling yields an exposure age of $25.4 \pm 3.1 \text{ ka}$, compatible with the Last Glacial
297 Maximum, and requires relatively low rates of denudation (about $20.9 \pm 11.2 \text{ mm/ka}$), (Table
298 2). These outputs include the uncertainties associated to the production rate, as they were
299 propagated through the Bayesian inversion. Uncertainties resulting from the estimated
300 density have been addressed by modeling the profile using different proportions of
301 pebble/matrix (Table 2).

302

303 **4. Glacier reconstruction, ELA, and paleo-climatic constraints**

304 The geomorphological evidence presented in section 2, as well as the LGM age control
305 on the emplacement of the diamicton at Manhale (section 3), collectively support the
306 hypothesis that the landform successions observed at the three sites are largely derived from
307 the action of former glaciers through abrasional, depositional, and melting processes. At
308 Terkmen and Qandîl, these glaciers could not have been much larger than $1.0\text{-}1.5 \text{ km}^2$, as
309 testified by the limited upstream glacial cirques catchments (Figures 2c and 3b). It thus
310 requires the vertical extension of the glaciers to be small, picturing them as small ice caps
311 developed from areas of snow accumulation on-the-spot. Therefore, the 2300-2400-m-
312 elevation of these two small cirques (Figure 2b and 3a) should provide a reasonable estimate
313 of the former local ELA and hence strongly suggests that these two glaciers were active
314 broadly at the same time.

315 On the other hand, the Manhale morainic ridges clearly lay at lower elevation than
316 those at Terkmen and Qandîl (about 300-350-m below), but must have had a much larger
317 upstream catchment (about 9.15 km^2). This is because it had to include the U-shape Sawda
318 valley, suspended on top of the Cedars cirque and stretching up to about 3000 m (Figures 2b
319 and 2c). Therefore, the 2000-m-elevation of the Manhale moraines yields only a minimum

320 estimate of the LGM ELA, since the downward extent of a glacier below the ELA is likely to
321 scale with the accumulation area (i.e. the glacier area above the ELA) (Pellitero et al., 2015).

322 We have quantitatively tested these first-order ELA estimates for the three glaciers by
323 combining the *GlaRe* (Pellitero et al., 2016) and *ELA calculation* (Pellitero et al., 2015)
324 toolboxes (both of them running in ArcGIS software). The equilibrium profile of each paleo-
325 glacier was first reconstructed based on their inferred geometries (shown in Figure 6a) using
326 *GlaRe*. The geometry of the Manhale glacier was inferred from both the trimlines of the Sawda
327 valley and the position and extent of the frontal moraines. The main uncertainty associated
328 with this reconstruction is on the left bank of the Sawda valley, where the trimline is not
329 continuous, and sometimes divides into two sub-parallel branches separated from each other
330 by up to 300m. In the *GlaRe* simulations, two distinct geometries of the paleo-glacier were
331 thus used to reconstruct the equilibrium profile (*smaller* and *larger* glacier; see solid and
332 dashed lines in Figure 6a). Similarly, two extreme geometries were used for the Terkmen and
333 Qandîl paleo-glaciers, one restricted to the inferred pro-glacial lake, and the other extending
334 up to the southeastern limit of the Cedars cirque (Figure 6a).

335 The glacier surface outputs yielded by the *GlaRe* simulations were then used to
336 calculate the corresponding ELA in the *ELA calculation* toolbox using the widely used
337 Accumulation Area Ratio (AAR) method (Pellitero et al., 2015). We considered AAR ratios
338 ranging between 0.65 and 0.7, thereby bracketing the widely accepted value of 0.67 (Pellitero
339 et al., 2015; see also Kern and Laszlo, 2010). The obtained ELAs range between 2410 and
340 2430m when considering the Qandîl ridge, 2395 and 2425m when considering the Terkmen
341 morainic ridge, and 2310 and 2650m when considering the Manhale morainic ridges. As
342 expected, the ELA of the Terkmen and Qandîl paleo-glaciers is close to the elevation of the
343 cirques, and this is irrespective of the chosen glacier geometries, as the *smaller glacier* and
344 *larger glacier* end-members yield ELAs that differ by only 20-30 m. In addition, the ELAs of the
345 Terkmen and Qandîl paleo-glaciers are undistinguishable from each other within uncertainties
346 (about 2400 m), which strongly supports the view that the two groups of landforms record a
347 single glacial event. The ELA obtained for the *smaller* Manhale paleo-glacier (2310-2430 m) is
348 also compatible with the 2400 m ELA estimated for the other two. Even the *larger glacier* end-
349 member with slightly higher values (2390-2650 m) is consistent within uncertainties. The
350 internal consistency between the ELAs obtained from the three paleo-glaciers within a small
351 geographic area strongly suggests that they all record a single glacial event, associated with

352 an ELA at about 2400m (i.e. the *smaller glacier* end-member is preferred over the *larger*
353 *glacier* end-member for the Manhale paleo-glacier).

354 The obtained ELA interval was further used to constrain paleo-climatic parameters.
355 Specifically, the glacier mass balance at the reconstructed ELA (2310-2650m) was computed
356 for various changes in precipitation and temperature relative to modern conditions, and using
357 a degree-day model. Degree-day models are based on the assumption that when the air
358 temperature is above the melting point, the ablation rate of a glacier is proportional to the
359 temperature (Braithwaite and Olesen, 1989). Accumulation is obtained as the product of
360 monthly precipitation and monthly probability of freezing, with the latter estimated from the
361 assumption that temperatures within a month are normally distributed with a standard
362 deviation of 4°C (Braithwaite, 1984). Yearly ablation is obtained as the product of the positive
363 degree-day sum and the degree-day factor, where the positive degree-day sum is the sum of
364 positive air temperatures (Braithwaite and Zhang, 2000) calculated assuming a sinusoidal
365 temperature variation throughout the year (Braithwaite et al., 2006; Brugger, 2006; Hughes
366 and Braithwaite, 2008). The degree-day factor is a quantity that describes how fast the melting
367 of snow (or ice) occurs along with increasing temperature above the melting point
368 (Braithwaite and Olesen, 1989).

369 Our model used modern climate conditions (monthly temperature and precipitation)
370 taken from an observation time range of 21 years between 1949 and 1970 (Plassard, 1981) at
371 Bcharre (1430m a.s.l.) located 5-6 km away from the paleo-glaciers ELA. Temperatures were
372 scaled to the altitude of the glaciers using the standard atmospheric lapse-rate (6.5°C km⁻¹)
373 (Barry and Chorley, 2019). Monthly precipitation was increased by a factor of 1.10 to 1.15
374 relative to the Bcharre data, in order to account for the 100 to 150mm difference in the annual
375 precipitation observed between Bcharre and the area of the reconstructed paleo-glaciers
376 (Figure 2a). The explored space of paleoclimatic conditions includes a range of temperature
377 cooling (T) between 1 and 14°C, and a factor of precipitation change relative to present (FP)
378 varying from x0.1 to x3.0. We applied a degree-day model for snow (most appropriate for
379 exploring conditions at the ELA; Hughes and Braithwaite, 2008; Braithwaite and Raper, 2007)
380 to these paleoclimatic conditions using an empirically constrained conservative interval of
381 4.1±1.5mm/day/°C for the degree-day factor (Braithwaite, 2008). We have neglected
382 accumulation through refreezing of melt water, which is justified by the fact that the annual
383 temperature at the reconstructed ELA is rather high (Braithwaite, et al., 2006): +2.5°C today

384 (and about -8°C at the time of moraine emplacement, see section 5.2). The zero-mass-balance
385 contours derived from both the preferred (2400m) (thick black curve) and extreme bounds
386 (2310 and 2650m) (curves bounding dark gray envelope) of the ELA, and for the preferred
387 degree-day factor of $4.1\text{mm/day}/^{\circ}\text{C}$, are provided in the (T;FP) plot of Figure 6b. The curves
388 bounding the light gray envelope consider the extreme bounds of the degree-day factor. For
389 each tested reconstruction (i.e. a given ELA associated to a given value of the degree-day
390 factor), the corresponding curve captures the space of (T;FP) solution needed to sustain the
391 reconstructed glacier. Paleo-climatic implications will be discussed in the next section.

392

393 **5. Discussion**

394 5.1. Presence of LGM glaciers on Mount Lebanon

395 The presence of LGM glaciers on Mount Lebanon is demonstrated by several lines of
396 arguments. First, prominent abrasional landforms are observed, such as the U-shape Sawda
397 valley and the over-deepening associated with former pro-glacial lakes at Terkmen and Qandil.
398 Second, typical landform successions suggestive of melting/depositional processes are
399 observed at the three studied localities (abrasional hollows, depositional or abrasional slope-
400 orthogonal ridges, and fields of hummocks), and culminate with more than 10-m-thick
401 diamicton deposit that forms the Manhale morainic ridges. Third, ^{36}Cl dating of the Manhale
402 diamicton demonstrates that these ridges were emplaced in a time interval consistent with
403 the Last Glacial Maximum. Fourth, the Equilibrium Line Altitude required to sustain these
404 paleo-glaciers (about 2400m) matches with the LGM ELA reconstructions at similar latitudes
405 in the Mediterranean (Figure 1a).

406 Even if diamicton could also be associated to landslide sediments, the age of the
407 deposit (LGM), and its spatial restriction to specific sites in the absence of fresh scars on the
408 walls of the Cedars cirque rule out a non-glacial origin for the ridge deposits observed at
409 Manhale. In addition, the Manhale group of ridges has an oblique linear trend relative to the
410 local orientation of the cirque wall (Figure 2c), which would not be expected if they had
411 resulted from landsliding. In addition, it is worth noting that the glacial landforms are
412 restricted to the W-facing wall of the Cedars cirque (Figure 2c), suggesting that the causative
413 agent of erosion and sedimentation did not affect the rest of the cirque, which has yet a similar
414 geology and morphology. As depicted in Figure 6a, this observation is strongly consistent with
415 the reconstructed ELA, since only the eastern part of the Cedars cirque reaches elevations that

416 are significantly above the 2400-m-contour, a condition necessary for snow accumulation
417 from year to year, and glacier formation. Another possible explanation might be that the ELA
418 was locally reduced on the W-facing side of the cirque because of higher rates of snow
419 accumulation, resulting from the effects of the dominant winds blowing eastward from the
420 Mediterranean (Maheras, 1988). In the latter case, the inferred 2400-m-ELA should be only
421 used with caution when deriving paleo-temperature and paleo-precipitation. Specifically, for
422 a given glacier extension and for a fixed temperature, the equilibrium precipitation should be
423 reduced relative to the estimates derived in section 4, to an extent that would depend on the
424 fraction of excess snow brought by wind transport from external sources (Hughes et al., 2006;
425 Styllas et al., 2018).

426 The relatively smooth landscape of the Mnaitra plateau and Sawda valley at elevations
427 well above the reconstructed ELA might seem surprising. We think that this particular
428 morphology has resulted from the combined effect of wind and karst. The exposition of the
429 plateau to dominant winds might indeed have hampered the accumulation of ice as already
430 proposed by Messerli (1966). Accordingly, the leeward Sawda valley would have been the only
431 portion of the plateau that allowed the development of a small glacier. Although the smooth
432 morphology of the Sawda valley is somewhat atypical (subdued trimlines in particular), it is
433 consistent with the landscapes observed in areas that underwent the combined erosional
434 effects of karstic and glacial processes. For example, large areas of the former Mount Orjen
435 ice-caps (Montenegro) (Hughes et al., 2010) now exhibit a very smooth morphology which is
436 much more typical of karst erosion than glacial erosion. Similarly, in the Northern Calcareous
437 Alps (Austria), smooth karstic landforms of the Hochschwab plateau are preserved on top of
438 deeply incised glacial valleys. Overall, we note that the morphology of our studied site bears
439 strong similarities with the western flank of the Crvanj mountains (Bosnia and Herzegovina),
440 which is topped by a gently dipping karstic plateau analogous to the Mnaitra plateau, and
441 covered by, presumably LGM, glacial moraines at its base (Zebre et al., 2019). In addition, our
442 reconstruction of the paleo-glaciers involves the Sawda valley glacial tongue, which extended
443 over a very gently sloping (less than 1°) trough between about 2950 and 2800m a.s.l. This
444 glacial tongue was irregularly connected to the ice mass that built the Manhale moraine
445 through a dramatically thinned section of ice cascading on the flank of the Cedars cirque
446 (Figure 6a). In that context, we think that the glacial tongue of the Sawda valley essentially

447 behaved as a plateau glacier, often not associated with the sharp morphologies typical of
448 glacial valleys (i.e. trimlines).

449 The uncertainties associated with the exposure ages obtained from the Bayesian
450 inversion of the ^{36}Cl Manhale depth-profile correspond to the 2σ Bayesian uncertainty
451 calculated from the sampling distribution of the Markov chain, and thus include the
452 uncertainties associated with both the ^{36}Cl measurements and the ^{36}Cl production rate. The
453 major additional source of uncertainty arises from the choice of the model bulk density.
454 Propagation of uncertainties associated with our density estimate into age uncertainties can
455 be assessed from Table 2. According to field observations and measurements, we consider
456 that the 1.6-1.8 g cm⁻³ interval, for which the Bayesian inversion yielded exposure-ages
457 ranging from 25.4±3.1 ka to 31.2±4.5 ka (Table 2), represents a conservative interval. The use
458 of the depth profile for those type of morainic deposits avoids problems related to erosion
459 and weathering being especially difficult to account for in the age estimates when sampling
460 solely surface samples (e.g. Žebre et al. 2019). These results place the maximum advance of
461 the Manhale glacier near the upper limit of the global LGM (26.5-19.0ka in Clark et al. (2009)
462 and 27.5-23.4 ka in Hughes and Gibbard (2015)). The LGM paleo-glaciers identified in the
463 present study represent the first evidence of LGM glaciers in the SE Mediterranean following
464 initial suggestion by Messerli (1966) (Figure 1a). The early timing of maximum glacier
465 extension in this part of the Mediterranean appears consistent with results obtained from the
466 closest glacial deposits preserved in climatically similar environments, i.e. in the Taurus range
467 of Turkey, and farther in Greece on the topographic highs of the Pindus Mountains and
468 Peloponnese. In these mountains, LGM glaciers generally advanced and stabilized before or
469 in the early phase of the global LGM: at about 30 ka on the Dedegöl Mountains (Köse et al.,
470 2019), at 35-28 ka on Mt Akdag (Sarıkaya et al., 2014), at 29-26 ka on Mt Tymphi (Allard et al.,
471 2020), and at 27-23 ka on Mt Chelmos (Pope et al., 2017).

472 In addition, the ELA reconstruction of the Mount Lebanon glaciers (~2400 m)
473 compares well with values obtained for the above-cited regions: between 2000 and 2700m in
474 Turkey with most LGM ELAs clustering around 2300-2500m (see the compilation of Sarıkaya
475 and Ciner, 2015), and between 2000 and 2100 m in Greece (Allard et al., 2020; Pope et al.,
476 2017). More generally, it is consistent with a LGM drop of the ELA to about 2400 m in the
477 transitional zone between the northern temperate climate and the sub-tropical deserts in the
478 eastern Mediterranean (between 34°N and 36°N), as suggested by the interpolation of

479 Mediterranean data (Figure 1a). ELA contours of the Levant region in Figure 1a were drawn
480 from the map of Messerli (1966), itself based on moraines identified at 2500m by the author.
481 Therefore, the lower ELA suggested by our results is simply a consequence of the slightly lower
482 elevation of the moraines reported in the present study. It thus suggests that the moraines of
483 Messerli (1966) might either be associated with slightly less extensive (younger?)
484 development of glaciers, or reflect slightly different local conditions.

485 We stress that, to a first order, the widespread absence of LGM glaciers in the SE
486 Mediterranean reflects the absence of high topography (Figure 1a). In that sense, the
487 presence of past glaciers in Lebanon appears directly related to the structural transpressive
488 bend along the Dead-Sea transform plate boundary (Figure 1a), across which Mount Lebanon
489 has been uplifted over the last few million years (Elias et al., 2007).

490

491 5.2. Local and regional paleoclimatic implications

492 The ELA interval constrained from the reconstruction of the Mount Lebanon LGM
493 glaciers has been converted into a solution space of cooling (ΔT) and factor of precipitation
494 change (FP) using a degree-day model (Brugger, 2006; Hughes and Braithwaite, 2008) (section
495 4 and Figure 6b). The thick black line in Figure 6b corresponds to our preferred estimate, which
496 is based on a 2400-m-ELA, a 100-mm-increase correction of the annual precipitation from the
497 climatic station of Bcharre to the area of the paleo-glaciers (based on the precipitation map
498 of Figure 2a), and a degree-day factor of 4.1mm/day/°C. The dark gray envelope encompasses
499 the conservative estimates, which allow the LGM ELA to range between 2310 and 2650m, and
500 the annual precipitation correction to reach up to 150 mm. As commonly observed in such
501 (ΔT ; FP) plots, the ELA is very sensitive to changes in temperature, but much less to changes
502 in precipitation (Figure 6b). For unchanged precipitation relative to the present, the ELA
503 interval constrained for the Mount Lebanon glaciers would indicate a cooling of ~7-8°C relative
504 to modern conditions (Figure 6b). On the other hand, the (ΔT ; FP) solution demonstrates that
505 the reconstructed glaciers would allow a maximum cooling of 10-11°C, a value obtained by
506 considering extremely dry, probably unrealistic, conditions (i.e. factor of precipitation change
507 reduced to 0.1) (Figure 6b). We recall that the possibility of excess snow brought by the
508 dominant winds blowing from the West, would imply that any reconstructed precipitation
509 represents a maximum precipitation (see section 5.1).

510 The major uncertainty associated with the LGM climate in the Levant region resides in
511 the hydrological regime: was the LGM climate dry or wet relative to the present? To address
512 this question, we summarize LGM cooling estimates in the Levant, and incorporate them into
513 the (ΔT ;FP) plot of Figure 6b to bracket possible values of FP. The considered paleo-
514 temperature records are of four types: alkenone-derived Sea Surface Temperature offshore
515 Israel (Almogi-Labin et al., 2009; Essallami et al., 2007), a pollen sequence in the Ghab valley,
516 northwest Syria (Wu et al., 2007), δD values of speleothem fluid inclusions in the Peqin cave,
517 northern Israel, and in the Soreq cave, Central Israel (McGarry et al., 2004), and clumped-
518 isotope thermometry on speleothem carbonates in the Soreq cave (Affek et al., 2008). The
519 LGM cooling determined by these studies range between 6°C and 14°C (in good agreement
520 with our estimate of 10-11°C at most) and are reported in Figure 6b. It has to be noticed that
521 these paleo-temperature records cover different periods of the LGM from 18.6 to 29.0 ka
522 (color-coded as “Late LGM” or “LGM” in Figure 6b).

523 Each of the paleo-temperature estimates has been converted into a factor of
524 precipitation change according to the space of (ΔT ; FP) solutions constrained by our glacial
525 reconstruction (double arrows in Figure 6b). The results show that all the continental paleo-
526 temperature records older than 20 ka invariably require a decrease in the rate of precipitation
527 when considering the classical value of 4.1mm/day/°C for the degree-day factor (orange
528 arrows in Figure 6b). We still note that the reconstructed glacier extent might be compatible
529 with a slight increase in precipitation under the condition that the degree-day factor
530 approaches the extreme bound of its conservative interval, that is close to 5.6mm/day/°C. This
531 possibility is very unlikely however given the distribution of degree-day factors derived from
532 direct observations (Braithwaite, 2008). Although ^{36}Cl dating indicates that the Manhale
533 glacier reached its maximum advance prior to 20 ka, we have derived FP values using Late
534 LGM paleo-temperature records (yellow arrows in Figure 6b) for comparison. It shows that
535 these late terrestrial records would be compatible with a wide range of FP, because the slightly
536 lower cooling reported for that time interval project onto the steeper (left-hand side) section
537 of the (ΔT ,FP) plot, where the ELA varies very little as a function of precipitation rates (Figure
538 6b). For the same reason, the marine records, which span the 19.0-24.0 ka interval, would
539 imply either a decrease (Almogi-Labin et al., 2009) or an increase (Essallami et al., 2007) in the
540 rate of precipitation. Given both the Early LGM age of the Manhale moraine and its terrestrial
541 location, the cooling of 8-11°C at Peqin cave (McGarry et al., 2004), of 8-14°C in the Ghab

542 valley (Wu et al., 2007), and of 9-14°C at Soreq cave (McGarry et al., 2004) appear as the most
543 appropriate paleo-temperature records to derive paleo-precipitation estimates from the
544 reconstructed ELA (orange arrows in Figure 6b). These records suggest that the local climate
545 must have been drier than the modern conditions to restrict the Mount Lebanon glaciers to
546 their reconstructed extents.

547 Moreover, it should be noted that the LGM cooling values used in Figure 6b are
548 constrained from sites lying at least 1500 m below the Mount Lebanon paleo-glaciers.
549 Therefore, if the LGM lapse-rate in the Mediterranean was steeper than 6.5°C/km as
550 suggested by Kuhlemann et al. (2008), then the cooling values used in Figure 6b would
551 underestimate the LGM cooling at our studied site. This would in turn enhance the
552 precipitation reduction implied by the ($\Delta T, FP$) plot. The same shift towards even drier
553 conditions would be obtained if the reconstructed glaciers were partly built from wind-blown
554 snow as discussed in section 5.1. A dry LGM over Mount Lebanon is consistent with the oxygen
555 isotope record of ostracod shells in the Yammouneh paleo-lake, located as close as 12 km
556 from our studied site. This indicates that the LGM was a period of local water deficit (Develle
557 et al., 2010). Similarly, dry conditions appear to characterize the last four glacial stages at the
558 same site (Gasse et al., 2015). The Jeita cave isotope record (from the Lebanon coast) (Cheng
559 et al., 2015) follows the same trend during LGM as the speleothem records further south
560 (Central and South Israel) (Bar-Matthews et al., 2003). However, the slow deposition of the
561 Jeita cave speleothem during the glacial period as well as the consideration of the source
562 effect from the Mediterranean Sea on the isotope record support less wet conditions during
563 most of last glacial in the region (Bar Matthews et al., 2019). LGM lake high-stands in the Lisan-
564 Dead Sea Basin, some 150 km further S in Israel (Figure 1b), have long been attributed to
565 regionally increased precipitation (Stein et al., 2010; Torfstein et al., 2013a and 2013b). A
566 recent study by Miebach et al. (2019) conducted on pollens from Lake Lisan sediments
567 provides evidence that the glacial pollen record was not necessarily associated with
568 precipitation increase during LGM. This is in line with a modeling study suggesting the glacial
569 high-stands in Lake Lisan were best explained by a decrease in evaporation that compensated
570 the concomitant decrease in the rates of precipitation (Stockhecke et al., 2016).

571 On a broader geographical scale, long-term transient simulations from a coupled
572 atmosphere-ocean-sea-ice-vegetation model (LOVECLIM) show that the northern hemisphere
573 ice sheets during glacial stages had a strong impact on atmospheric circulation causing a

574 massive drying in the eastern Mediterranean and in the Levant (Stockhecke et al., 2016). As
575 shown by a LGM simulation from a general circulation model (PMIP-2), the mid-latitude
576 westerly belt was pushed southward in response to the ice sheet and sea ice extent at
577 northern high latitudes (Lai  n   et al., 2009).

578

579

580 **6. Conclusions**

581 We have identified morphological and sedimentological evidence that testify to the
582 existence of past glaciers on the western side of Mount Lebanon. The maximum advance of
583 three glaciers (one 8-km-long glacial tongue and two small glacier cirques) has been inferred
584 from their geometries. The deposition of the largest glacier frontal moraine has been dated
585 to 25.4 ± 3.1 - 31.2 ± 4.5 ka based on ^{36}Cl Cosmic Ray Exposure dating of a 10 m deep profile in
586 the constitutive diamicton. The reconstructed glacier hypsometry indicates that this
587 maximum glacier extension occurred along an Equilibrium Line Altitude depressed to about
588 2400 m using the AAR method. The early LGM age obtained for the maximum glacier advance
589 matches with glacial chronologies obtained from other Eastern Mediterranean moraines (in
590 particular in Turkey and Greece). Finally, a degree-day model was used to infer the LGM paleo-
591 climatic conditions required to depress the ELA to about 2400 m. It suggests that, under the
592 LGM cooling conditions documented for the Levant region, the local climate was drier
593 compared to modern conditions. This study highlights the complexity of Mediterranean
594 climatic responses, in an area influenced by several factors (large scale climate pattern, local
595 orography, active tectonics). Inputs of several climate archives, although influenced by
596 different biases, provide clues to answer the questions of spatial heterogeneities of the
597 palaeohydrological response in the Levant.

598 In the Mediterranean region it is therefore crucial to combine all approaches to
599 reconstructing climate change in order to resolve fine-scale variations and to validate certain
600 paleoclimatic proxies.

601

602

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609

610

611 **7. References**

612

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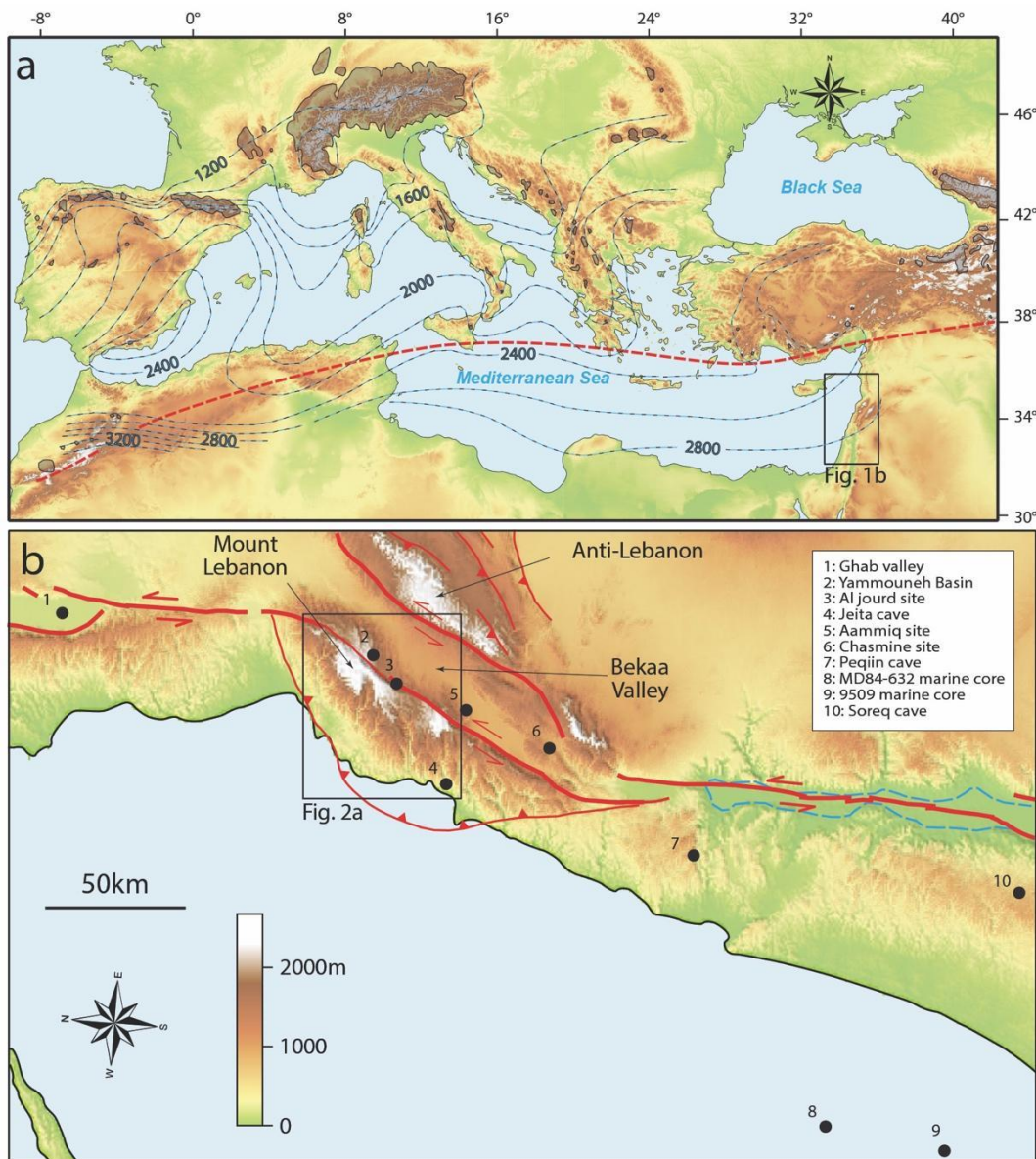
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894 **Figures and Tables**

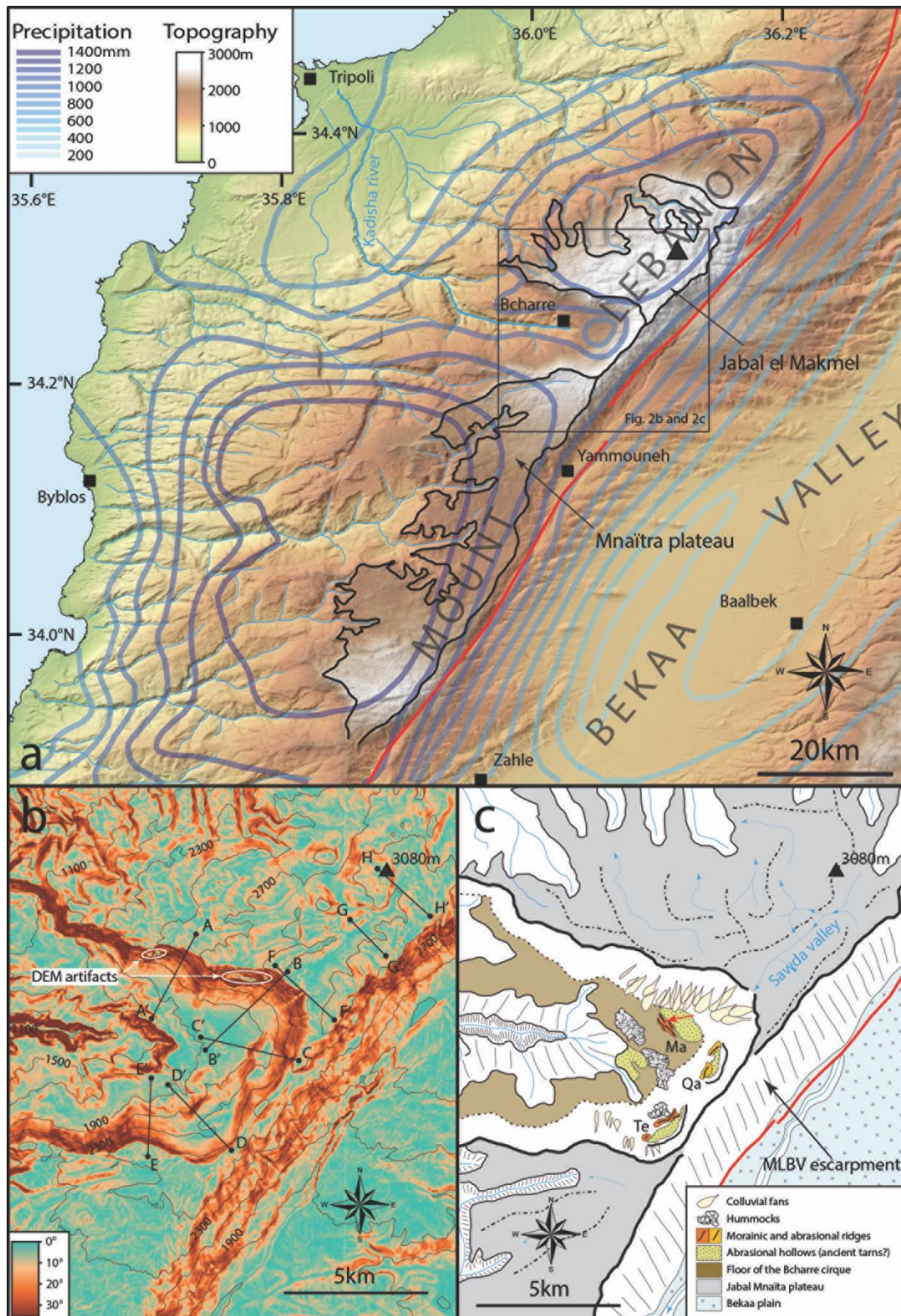


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896 *Figure 1: Location of the study site in the context of previous paleo-climate records in the Mediterranean (a)*
897 *Interpolated ELA (dashed black-blue lines) constructed based on local maximum advances of glaciers around*
898 *the Mediterranean basin (Kuhlemann et al. (2008) completed by map of Messerli (1966) for the southeastern*
899 *corner of the Mediterranean). Note that the highlighted ELA pattern is thus diachronic since some*
900 *Mediterranean glaciers reached their maximum in the LGM (e.g. Turkey) while some other did in the Middle*

901 *Pleistocene (Balkans, Greece). However, it still provides a broad view of the atmospheric structure during glacial*
902 *conditions. The extent of LGM glaciers is shown in transparent black (after Allen et al., 2008; Ehlers and*
903 *Gibbard, 2004; Kuhlemann et al., 2008). The dashed red line represents the southern limit of known LGM*
904 *glaciers in the Mediterranean. (b) Shaded relief of the Levant region showing the location of the main paleo-*
905 *climatic records discussed in the text. The main active faults are drawn in red after Daëron et al. (2007). The*
906 *blue dashed line represents the extent of paleo-Lake Lisan. Topographic scale is the same in (a) and (b).*

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908

909 *Figure 2: Location of the study site in the context of local precipitation, topography and geomorphology. (a):*
 910 *Drainage network around Jabal el Makmel. Catchment limits are depicted with dashed black lines. The solid black*
 911 *line delineates Mt-Lebanon plateau. Precipitation contours are same as in Figure 1b (Cheddadi and Khater (2016)*
 912 *after Plassard (1981)). Black triangles denotes the Qurnat as Sawda', the highest peak of Mount Lebanon*
 913 *(3088m). (b) Slope map of the Cedars cirque and surrounding areas, derived from a 20-m-DEM of Lebanon (i.e.*

914 SPOT DEM product extracted from SPOT 5 HRS stereoscopic data). 200-m interval elevation contours are
915 superposed over the slope map. Black lines show location of transects of Figure 4. (c) Geomorphic interpretation
916 of the same area as in (b) based on field observations and the analysis of 2.5-m-SPOT imagery, 1:20000
917 topographic maps, and the 20-m DEM of Lebanon. Ma=Manhale, Qa=Qandil, Te=Terkmen. MLBV
918 escarpment=MountLebanon-BekaaValley escarpment. Yellow versus orange ridges refer to the abrasional versus
919 depositional nature of the mapped ridges. Red arrow locates the ³⁶Cl depth-profile. The internal topography of
920 the plateau is represented by the watershed ridges (thick black dashed lines) and thalwegs (dashed blue arrows).

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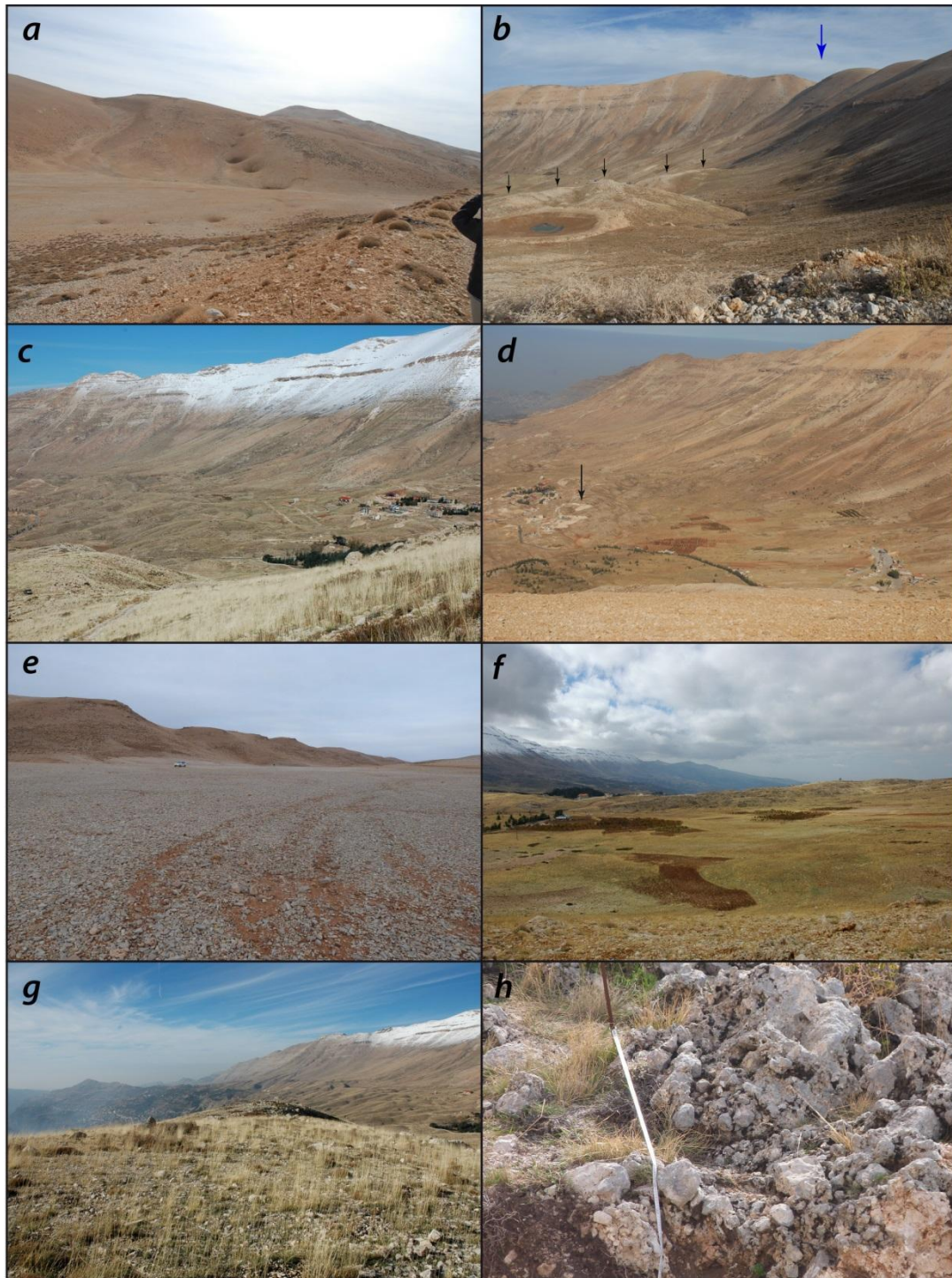
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928 *Figure 3: Field pictures: (a) SE view of the Sawda valley (note the presence of sinkholes at the base of the slopes),*
 929 *(b) the Qandil ridge (black arrows) and the northern and eastern walls of the Cedars cirque in the background*
 930 *(looking N). The blue arrow shows the outlet of the Sawda valley, (c) view looking NNW showing the fields of*
 931 *hummocks in Manhale and the northern wall of the Cedars cirque in the background, (d) the Manhale ridges*
 932 *(black arrow shows the location of the excavation where the ³⁶Cl depth profile was sampled) looking WNW, and*
 933 *the cultivated clay-rich depression on the back, (e) the flat bottom of the Sawda valley floored by sorted (non-*
 934 *patterned) clasts, (f) WSW view with the Manhale ridges (in the middle-background) and its associated clay-rich*

935 *depression (in the foreground), (g) the Turkmen morainic ridge looking NW, (h) the morainic deposit cemented*
936 *by secondary calcite on top of the ³⁶Cl depth-profile of Figure 5.*

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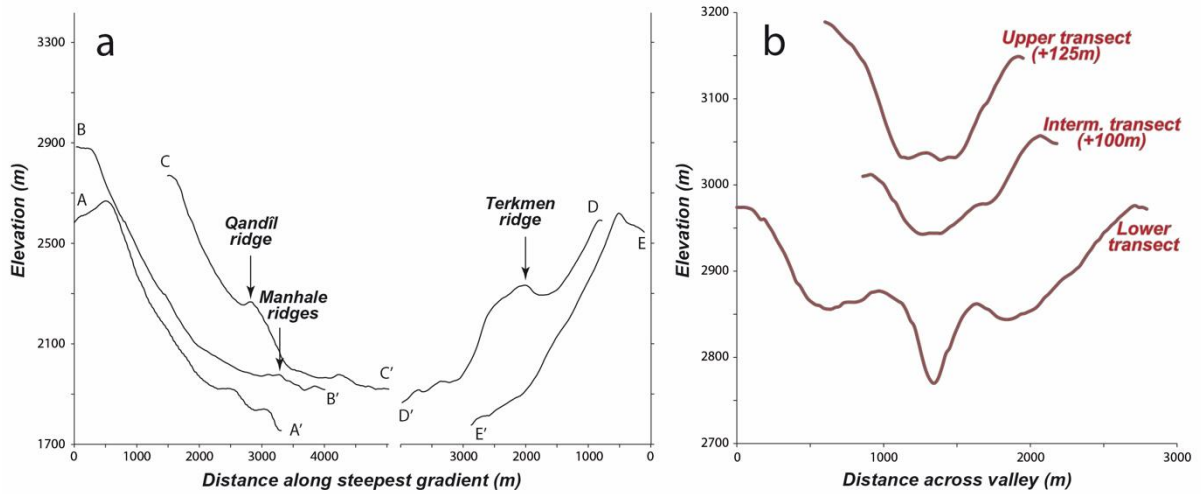
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955 *Figure 4: Topographic profiles (extracted from the 20-m-DEM of Lebanon; see Figure 3a for location) across the*
 956 *walls of the Cedars cirque (a) and across the Sawda valley (b). Note the abrasional hollows behind the Qandil and*
 957 *Terkmen ridges (profiles DD' and CC' in (a)), and the U shape of the Sawda valley (b). Values in brackets in (b)*
 958 *indicate the magnitude of upward shift which has been applied to the intermediate and upper profiles for the*
 959 *sake of clarity.*

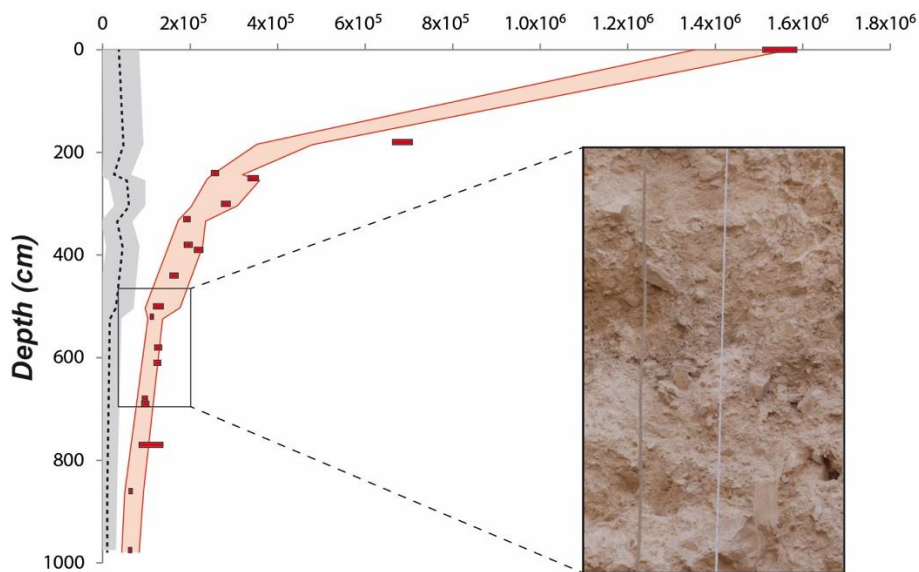
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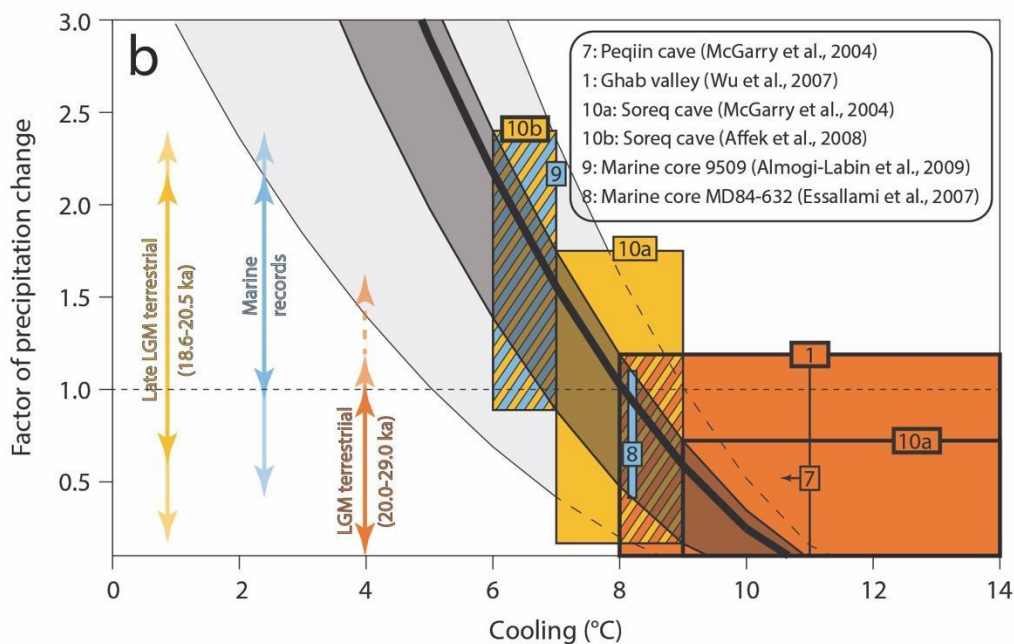
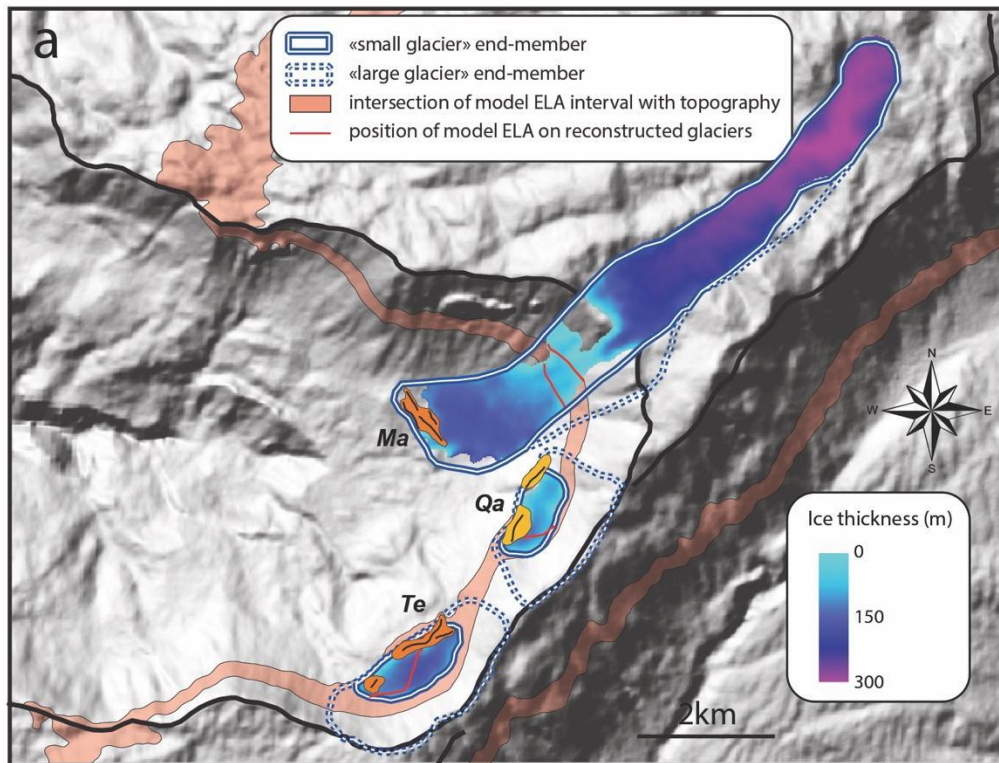


^{36}Cl concentration (at/g of rock)



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964 *Figure 5:* (Above) SE view of the Manhole moraine excavation site wall, with position of collected samples
 965 indicated by red arrows. (Below) The graph of measured ^{36}Cl concentrations plotted against depth of the
 966 corresponding sample using same vertical scale as the field picture. Red boxes are for sample's concentration
 967 where box width relates to the 1σ uncertainty. Red envelope represents the 1.6-density model output including
 968 2σ uncertainties. The breaks in the exponential trend reflect the effect of sample-specific concentrations of target
 969 elements. Black dashed curve and gray envelope represent the sample-specific inheritance and associated 2σ
 970 uncertainties yielded by the Monte Carlo inversion. The inset picture is a close-up of the diamicton deposit
 971 between 5m and 7m depth.



972

973 *Figure 6: ELA reconstructions and paleoclimatic implications. (a) The results of glacier and ELA reconstructions*
 974 *are summarized onto the shaded relief of the Cedars cirque. The ice thickness of the three glaciers comes from*
 975 *the simulations performed by the GlaRe toolbox of Pellitero et al. (2016). Ma=Manhale, Qa=Qandil, Te=Terkmen.*
 976 *(b) (ΔT ;FP) plot. The transparent black envelope represents the changes in precipitation and temperature relative*
 977 *to modern conditions required to depress the LGM ELA into the conservative interval constrained from our*
 978 *simulations (as obtained from a degree-day model). Gray envelope is the same but also includes sensitivity to*
 979 *variations in degree-day factor of snow (from 2.6 to 5.6mm/d/K). The thick black line depicts the preferred ELA*

980 *(see text for more informations). The rectangular boxes have been constructed based on paleo-temperature*
981 *estimates of the Levant region (see inset for references) and their compatibility with the (ΔT ;FP) plot, and are*
982 *color-coded according to the type of records (orange: Early LGM terrestrial record; yellow: Late LGM terrestrial*
983 *record; blue: marine records). The double arrows denote the range of FP constrained by these boxes using either*
984 *the gray envelope (dashed transparent arrow, only shown for orange color code), or the black envelope*
985 *(transparent arrows), or the thick curve (non-transparent arrows). Hatched areas correspond to overlapping of*
986 *two types of records.*

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1012 *Table 1: Site description, general sample information, measurements ($^{36}\text{Cl}/^{35}\text{Cl}$ and $^{35}\text{Cl}/^{37}\text{Cl}$ ratios and chlorine*
 1013 *and CaO concentrations) and calculated ^{36}Cl and chlorine concentrations. Whole-rock composition can be*
 1014 *accessed through online Supplementary Table.*

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Elevation 2006	Latitude 34.24969	Longitude 36.0517	Shielding 0.986	^{36}Cl blank corr. $5.8 \cdot 10^4$ atoms					
Sample	depth (cm)	dissolved weight (g)	mass of Cl in spike (mg)	$^{36}\text{Cl}/^{35}\text{Cl}$ ($\times 10^{-14}$)	$^{35}\text{Cl}/^{37}\text{Cl}$	^{36}Cl ($\times 10^4$ at/g)	Cl (ppm)	CaO (%)	
MAP11-1	0	82.63	2.37	354.31±7.63	9.74±0.02	154.7±3.9	11	55.9	
MAP11-20	180	83.86	2.39	132.77±3.83	7.29±0.02	68.5±2.3	17	55.7	
MAP11-3	240	83.92	2.36	71.42±2.27	16.28±0.04	25.7±0.9	5	55.0	
MAP11-4	250	96.68	2.36	84.02±2.63	8.15±0.02	34.4±1.2	12	55.9	
MAP11-21	300	81.55	2.38	64.49±2.08	10.10±0.03	28.1±1.0	10	56.6	
MAP11-5	330	76.50	2.36	32.55±1.09	6.76±0.09	19.3±0.8	21	44.2	
MAP11-6	380	70.05	2.35	28.74±1.19	6.36±0.08	19.5±0.9	26	46.7	
MAP11-7	390	76.66	2.37	21.32±0.77	4.53±0.02	21.9±1.0	55	44.0	
MAP11-8	440	51.23	2.37	15.41±0.83	5.67±0.04	16.3±1.0	46	44.5	
MAP11-9	500	74.50	2.38	16.36±1.41	5.42±0.05	12.7±1.1	35	45.7	
MAP11-10	520	77.07	2.38	23.17±0.70	9.03±0.08	11.2±0.4	13	47.0	
MAP11-11	580	65.87	2.38	9.40±0.52	4.31±0.03	12.7±0.8	76	34.8	
MAP11-12	610	51.37	2.38	9.81±0.58	4.99±0.01	12.5±0.8	62	39.5	
MAP11-14	680	53.02	2.37	7.34±0.40	4.80±0.02	9.6±0.6	67	32.9	
MAP11-13	690	51.33	2.38	6.29±0.50	4.51±0.01	9.7±0.8	84	32.3	
MAP11-15	770	4.76	2.39	2.02±0.44	20.82±0.06	11.1±2.7	69	33.0	
MAP11-16	860	69.24	2.39	7.77±0.44	5.48±0.01	6.4±0.4	37	38.2	
MAP11-17	975	77.60	2.40	6.73±0.34	4.73±0.01	6.3±0.4	48	42.2	

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1022 *Table 2: Results of profile modeling using Bayesian inversion (3000000 iterations). RMS (Residual Mean Square)*
 1023 *shows no convergence towards reasonable density values, implying that model outputs cannot be used to*
 1024 *constrain the density.*

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density ($\text{g}\cdot\text{cm}^{-3}$)	exposure age (ka)	erosion-rate (mm/ka)	RMS
1.2	20.1±0.9	6.3±6.3	28
1.4	21.7±1.4	10.4±8.1	37
1.6	25.4±3.1	20.9±11.2	62
1.8	31.2±4.5	29.9±10.4	85
2.0	37.9±6.3	33.9±9.2	103

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