

Aeolian dust accretion outpaces erosion in the formation of Mediterranean alpine soils. New evidence from the periglacial zone of Mount Olympus, Greece

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Abstract:	<p>Soil formation in Mediterranean periglacial landscapes remains poorly understood as the interplay between erosion and aeolian dust accretion in providing parent materials, and mineral weathering and pedogenesis, as dominant post depositional processes, depends on a variety of local and regional factors. Herein, we investigate the balance between erosion and aeolian dust accretion in the formation of an alpine soil profile along an erosional gradient in the periglacial zone of Mount Olympus in Greece. We applied a wide range of analytical methods to 23 samples, from a soil profile developed in a glaciokarstic plateau, from colluvial sediment horizons interbedded in postglacial scree slopes and from modern Sahara dust samples deposited on the snowpack. Colluvial sediment horizons exhibit high concentrations of calcite rich sand and represent the local erosion products. The soil B horizon developed on a glaciokarstic plateau contains high amounts of fine earth and is rich in quartz, mica, plagioclase, clays, and Fe-Ti oxides. Based on its physical and textural characteristics the soil profile is partitioned in a surficial weathered Bw and a lower illuvial Bt horizon that overlies the local regolith composed of fragmented glacial till and slope wash deposits. Radiogenic isotope systematics, textural and mineralogical analysis show that the contribution of Sahara and locally sourced dust to the development of the soil B horizon ranges between 50 and 65%. Cryoturbation results in fine earth translocation from Bw to the Bt horizon, whereas weak pedogenetic modifications of aeolian and bedrock-derived minerals result in magnetic mineral weathering and secondary clay formation. Our findings reveal that, aeolian dust accretion is the dominant process in providing alpine soil parent material and that cryoturbation, weak pedogenesis, and clay mineral alteration occur within the Mediterranean periglacial zone of Mount Olympus.</p>

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2 **alpine soils. New evidence from the periglacial zone of Mount Olympus,**
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4

5 **Abstract**

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27 **Keywords:** alpine soil; erosion; aeolian dust accretion; mineral weathering; Mediterranean periglacial
28 zone, Mount Olympus

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31 1. INTRODUCTION

32 Global glacier retreat and the melting of permafrost and ground ice have altered the dynamics of the
33 alpine critical zone by enhancing erosion and by disturbing the production of mountain soils
34 (Haeberli et al., 2006, Egli et al., 2014). During periods of glacial retreat and paraglacial adjustment,
35 alpine soils develop from parent materials sourced through a combination of frost shattering,
36 colluvial activity, and hillslope outwash (Egli and Poullenard, 2016). An equally important factor that
37 affects the formation and evolution of alpine soils is the accretion of local and long-range
38 transported aeolian dust (Muhs and Benedict, 2006; Küfmann 2008; Lawrence et al., 2013; Drewnik
39 et al., 2014; Yang et al., 2016; Gild et al., 2018; Munroe et al., 2019). Thus, the contributions of
40 physical erosion and aeolian dust accretion are fundamental sources of alpine soil parent material
41 and largely define their textural, mineralogical, and geochemical characteristics.

42 The postglacial adjustment of alpine valleys is inherently linked to high rates of erosion, with
43 frequent rockfalls, debris flows, rock avalanches, and high rates of sediment production especially
44 below steep rockwalls. In such dynamic environments, alpine soil mantles formed on the surface of
45 slope deposits are patchy, often truncated and constantly rejuvenated from rockfall material,
46 whereas the evolution of these soils alternates between progressive and regressive phases (Egli et
47 al., 2018). Similar soil mantles developed on sandy layers deposited on the surface of stratified scree
48 slopes are generally indicative of quiescent periods of slope processes and are thus concise indicators
49 of optimum climatic conditions and alpine landscape stability (Sanders et al., 2010). When the
50 regional climate shifts to a colder regime, intense freeze–thaw activity and frost cracking enhance
51 rockfall activity and result in the erosion and gradual burial of these incipient soil mantles. As
52 hillslope processes and scree slope aggradation diminish away from the alpine steep rockwalls, the
53 development of alpine soils on distal moraines, outwash plains, and glacially scoured plateaus can be
54 considered continuous (*sensu lato*). In these depositional environments, low erosional rates provide
55 ample time for pedogenetic processes such as chemical weathering, mineral alteration, elemental
56 translocation, and illuviation to occur, whereas other physical processes, such as cryoturbation

57 disturb the soil profiles. Alpine soils are an important component of high mountain ecosystems, so a
58 better understanding of the processes that drive their formation in climatically sensitive regions, like
59 the Mediterranean, is required.

60 Most soils formed in the Mediterranean basin display a distinguishable red color (terra rossa)
61 that derives from high concentrations of ultra-fine pedogenic iron oxides, mainly hematite (Yaalon,
62 1997; Durn et al., 1999). Terra rossa soils receive significant aeolian dust additions from Sahara and
63 Sahel regions (Yaalon, 1997; Durn, 2003; Stuut et al., 2009). In the Mediterranean alpine hinterland,
64 thin drapes of Sahara-dust-rich soils are found on plateaus, glacial moraines, and outwash plains
65 (e.g., Rellini et al., 2009), whereas aeolian dust accretion in terra rossa soils can also originate from a
66 wide range of alluvial deposits, such as sand dunes, desiccated alluvial planes, and Quaternary loess
67 (Amit et al., 2020; Lehmkuhl et al., 2020). Most of the Mediterranean mountains are built up by
68 carbonate rocks, hence the aeolian input to alpine soil formation occurs in parallel with colluvial
69 deposition of carbonate erosion and dissolution products that form a characteristic insoluble residue
70 incorporated in the soil sequences (Durn 2003; Varga et al., 2016; Kirsten and Heinrich, 2022).

71 In the present study, we investigate the major processes that drive the postglacial formation of
72 Mediterranean alpine soils in the periglacial landscapes of Mount Olympus, Greece. We follow a
73 combined sedimentological, mineralogical, and isotopic approach, and present a detailed
74 characterization of distinct alpine sediment and soil horizons developed across a geomorphological
75 gradient of decreasing erosive power. Discrete sediment samples from intact sandy layers
76 interbedded in postglacial stratified scree slope deposits that represent *in situ* erosional products of
77 the periglacial zone of Mount Olympus, are compared with samples from a soil profile developed in a
78 glaciokarstic plateau, with a goal to assess the relative contributions of aeolian dust accretion to the
79 fine fraction of an alpine soil. We differentiate between the physical and chemical processes that
80 drive the production of the scree slope sandy layers and of the alpine soil profile, by comparing their
81 respective grain size distributions, and bulk mineralogy. Furthermore, we examine the potential
82 influence of Sahara and locally sourced aeolian dust accretion on the alpine soil profile by comparing

83 the sedimentological, mineralogical, and radiogenic isotope compositions through the application of
84 $^{86}\text{Sr}/^{87}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios between the soil samples and Sahara dust samples collected from the
85 snowpack. We finally measured the magnetic properties of the soil samples and clay mineralogy of
86 bottom and topsoil layers, to assess the potential for weathering of clay minerals and iron oxides
87 within Mount Olympus periglacial zone. Understanding the sources of parent materials and soil
88 formation processes between contrasting geomorphological settings is a fundamental step towards
89 defining the postglacial paleo-environmental history of Mount Olympus alpine landscapes that
90 followed pronounced shifts of the regional climate.

91

92 **2 BACKGROUND**

93 **2.1 Mount Olympus glacial history**

94 Mount Olympus is the highest mountain in Greece, rising 2918 m above the northwest coastline of
95 the Aegean Sea (Figure 1a). It is a precipitous massif with a circular shape composed of Triassic to
96 Cretaceous metacarbonates, uplifted along a frontal fault that runs parallel to the present-day
97 shoreline. Mount Olympus is exhumed from the silicate crystalline bedrock, which dominates the
98 lithology of Pieria Mountains (granites, ophiolites) to the north and east, and Mount Olympus
99 granites to the west (Figure 1B). High uplift rates along with successive Quaternary glaciations have
100 created the present-day rugged terrain. The deglaciation of Mount Olympus since the Last Glacial
101 Maximum (LGM), between 28 and 24 ka BP (Allard et al., 2020), triggered the rapid retreat of an ice
102 cap that was covering the summit area and extended down to elevations of ≈ 1800 m (Kuhlemann et
103 al., 2008). The post-LGM glacier retreat was intercepted by a glacier re-advance phase at ≈ 15 ka BP
104 that was limited at the highest cirques above 2200 m at ≈ 12.5 ka BP (Styllas et al., 2018). This latter
105 phase of glacial expanse is traced in both Megala Kazania (MK) and Throne of Zeus (TZ) cirques
106 (Figure 1C). The absence of absolutely dated glacial features between early- and mid-Holocene (9–4
107 ka BP) in both cirques suggests reduced glacial activity, whereas Late Holocene (4 ka BP to present)
108 glacier advances have been observed only in the MK cirque (Figure 1C). These include a terminal

109 moraine stabilization phase at \approx 2.5 ka BP followed by a smaller expansion of the MK glacier at the
110 beginning of the Little Ice Age (LIA) at \approx 0.6 ka BP (Styllas, 2020). Late Holocene glacier advances in
111 the MK cirque lack similarly dated glacial landforms in the TZ cirque, but we cannot rule out the
112 possibility that the Late Holocene climatic shifts towards glacial conditions triggered an
113 intensification of glacial and periglacial processes, which in turn affected the late Holocene landscape
114 evolution, scree slope aggradation and alpine soil production.

115

116 **2.2 Climate**

117 The contemporary maritime conditions and the steep relief of Mount Olympus result in intense
118 precipitation and temperature altitudinal gradients, with the highest peaks constituting an
119 orographic and climatic barrier between the eastern (marine) and western (continental) sides (Figure
120 1b, Styllas and Kaskaoutis, 2018). The climate in the coastal zone is typically Mediterranean, whereas
121 at higher elevations (1000–2200 m), the climate attains sub-Mediterranean characteristics with
122 average annual precipitation of 1300 mm (Styllas et al., 2016). In the alpine zone above the tree line
123 (2400 m), the climate is characterized by temperate conditions with annual precipitation above 2000
124 mm and average annual temperatures between 0 and 1.5 °C (Styllas et al., 2016). The periglacial
125 activity in the Mount Olympus alpine zone is likely still active today, as it is situated just above the
126 lower limit of the regional permafrost zone (2700m) of the southern Balkan peninsular (Dobiński,
127 2005).

128

129 **2.3 The Plateau of Muses**

130 The Plateau of Muses (PM) is a planar depositional surface located at an elevation of 2600 m with a
131 surface area of 1 km². It resembles a typical alpine meadow, partly covered by alpine grass
132 vegetation that shares similar characteristics with plateaus found in the high Balkan Mountains and

133 the European Alps. The PM is bounded to the south by the TZ cirque lateral moraine ridge and by
134 several gentle-sloping glacially eroded peaks along its northern, eastern, and western margins (Figure
135 1C). The formation of the plateau has resulted from the combined action of glacial scouring and
136 carbonate bedrock dissolution. Its low relief in combination with the circular shape suggest a doline
137 type karstic depression that is filled with glacial till, overlain by colluvial sediments (slope wash)
138 transported from the adjacent slopes. The surface layer of the PM sedimentary sequence comprises
139 a developed soil sequence with variable thickness (30–50 cm) that overlies a layer of outwash sand
140 and fine gravels and/or fragmented till boulders, and exhibits brown-red to yellow color hues, which
141 in the Munsell color scale range between 7.5 and 10 YR (Table 1). Alternating patches of alpine grass
142 vegetation and hummocky soil pans in the center of the plateau are indicative of periglacial activity
143 and cryoturbation. Other periglacial features such as solifluction-terraced stripes below the bare
144 bedrock of the surrounding summits are tentatively considered to have formed during the Late
145 Holocene cold stages, during the observed expansion of small glaciers in the MK cirque.

146

147 **3 MATERIALS AND METHODS**

148 **3.1 Erosional products and alpine soil sampling**

149 To adequately address the question of the relative balance between aeolian dust accretion and local
150 erosion of moraines and scree slopes to the development of the alpine soil on Mount Olympus
151 periglacial zone, a wide range of methods were employed and involved the analyses of 21 discrete
152 soil and sediment samples retrieved along a transect of decreasing hillslope energy and erosional
153 power (Figure 2). Five samples (n=5) were retrieved from clast-free sandy horizons interbedded in
154 the relatively young (Late Holocene) MK and older (early Holocene) TZ stratified scree slopes, and
155 sixteen (n=16) were sampled from the PM soil sequence at 2-cm intervals (Table 1). The specific
156 experimental setting was selected to evaluate the impact of physical weathering on providing the
157 base material for the development of the PM soil. We only sampled naturally exposed clast free
158 sandy layers found within the scree slopes of MK and TZ. We considered that these layers share

159 similar textural, mineralogical, and geochemical characteristics with the PM soil basal horizon, which
160 lies on a layer of outwash sand and gravels. Luckily, we were able to retrieve the samples from two
161 distinct interbedded clast free sediment layers within the TZ scree slope after a torrential rainfall
162 event that opened a deep erosional trench in the scree below the rockwall and reached the basal till
163 layer (Figure 2). The scree slope in the MK is regularly eroded and scoured from a perennial snowfield
164 that is retreating by the end of the summer season, and this made the sampling of distinct soil-
165 sediment horizons straightforward. We manually excavated only one pit for high resolution soil
166 sampling and considered that due to the very small surface area of the surficial soil apron within the
167 PM catchment (0.06 km²), the specific profile is representative of the PM soil development. We
168 selected a location in the center of a circular soil-sediment pan that was free of vegetation, surface
169 carbonate fragments (Figure 2). After sampling, the pit was closed and refilled with the excavated
170 material in accordance with Mount Olympus National Park directions. In locations with long lasting
171 snowpack, we observed a humic A horizon, but since these locations host several endemic flower
172 species, the Management Unit of Mount Olympus National Park, did not grant permission to
173 excavate a soil pit in these sensitive sites. The PM soil samples were additionally subjected to
174 microscopic and radiogenic isotope analyses and magnetic measurements to investigate the
175 potential chemical alterations processes during PM soil development. Mineralogical and radiogenic
176 isotope analyses were also performed in two (n = 2) samples of aeolian dust that were deposited on
177 the PM snowpack during the spring seasons of 2018 and 2022. The long-range aeolian dust transport
178 episodes occurred on March 22–24, 2018, and March 16–18, 2022. The synoptic conditions of these
179 distinct episodes show that the dust emissions traveled to Mount Olympus from the Sahara Desert
180 and left an orange hue on the snowpack, which later in the spring season formed distinct layers in
181 the snowpack (Figure 3). We therefore consider the samples collected from the PM snowpack as
182 representative of Sahara dust accretion in Mount Olympus alpine soils.

183

184 **3.2 Grain size analyses**

185 The soil samples were transported to the lab, wet sieved through a 3.5-mm sieve, and treated with
186 30% hydrogen peroxide (H_2O_2) at 70 °C for 12 h to remove organic matter. The H_2O_2 treatment was
187 repeated three times until the samples were completely bleached and all organic matter was
188 degraded. The samples were washed with distilled water and analyzed with a Mastersizer 3000 laser
189 diffraction particle-size analyzer to define the bulk grain size distributions of the sand, silt, and clay
190 fractions. The samples were run through the automated dispersion unit and sodium
191 hexametaphosphate solution (Calgon) was added as dispersion factor. Statistical analyses of the grain
192 size distributions and derivation of the clay, silt, and sand fractions were realized with MATLAB Curve
193 Fitting Lab (CFLab), which performs curve fitting on sediment grain size distributions using the
194 Weibull probability distribution function (Wu et al., 2020).

195

196 **3.3 Mineralogy**

197 Identification of the mineral phases of the soil and aeolian dust bulk samples was achieved through
198 X-ray diffraction (XRD, Philips diffractometer PW1800, Co radiation at 40 kV and 40 mA), and two
199 samples from the top and base of the PM soil profile (PM1 and PM15) were additionally analyzed for
200 their clay ($<2 \mu\text{m}$) mineralogy through ethylene glycolation and heating for 2 h at 550 °C. The PM soil
201 samples semi-quantitative composition of the main mineral phases (e.g., quartz, feldspar,
202 plagioclase, micas, calcite) was determined using MAUD-Material Analysis software applied for full
203 pattern Rietveld refinement (Lutterotti et al. 2007) and is expressed as weight percent (wt %)
204 concentrations.

205

206 **3.4 Petrographic, magnetic, and isotopic analyses**

207 Additional analytical methods were applied only to the PM soil samples to assess the potential sources
208 of soil-forming material, pedogenesis, and chemical weathering. The fabric configuration of the PM
209 alpine soil was explored through scanning electron microscopy–energy dispersive spectrometry
210 (SEM–EDS) analyses (JEOL JSM-840A equipped with an INCA 250; Oxford) with a 20-kV accelerating

211 voltage and 0.4-mA probe current. Backscattered electron images (BSE) enabled us to detect the
212 shapes of different minerals, and the physical weathering features of specific grains, whereas with
213 the EDS analysis we examined areas of different chemical composition within the same soil
214 aggregates.

215 We additionally explored the existence of ferromagnetic components and the potential for
216 secondary iron oxides formation in the PM soil profile through magnetic susceptibility
217 measurements. The discrete samples were packed in cubical plastic boxes ($2 \times 2 \times 2$ cm) and weighed
218 before the measurements. Volume-specific magnetic susceptibility measurements were performed
219 using both a Bartington dual MS2B sensor at low and high frequencies of 0.465 and 4.65 kHz. The
220 results are expressed as mass-specific magnetic susceptibility (χ ; 10^{-8} m³/kg). During the measuring
221 procedure, every sample was measured at least three times and the average value was assigned as
222 the final measurement. For each sample, two air measurements were performed before and after
223 sample measurement. The frequency-dependent susceptibility (χ_{FD} ; %) was calculated according to
224 Dearing et al. (1996):

$$225 \quad \chi_{FD}\% = \frac{100(\chi_{LF} - \chi_{HF})}{\chi_{LF}} \quad (1)$$

226 where χ_{LF} , χ_{HF} , are the magnetic susceptibility at low and high frequency, respectively. Samples PM16
227 and PM15, which were considered as more representative of the PM soil regolith boundary, were
228 additionally subjected to thermomagnetic analysis to define the origin of the ferromagnetic particles
229 at the base of the PM soil. Measurements of continuous thermomagnetic curves (K–T curves) at low
230 and high temperature were realized with the furnace CS3 of the AGICO MFK1-FA susceptibilimeter.

231 The potential sources of the PM soil and aeolian dust were evaluated through their Sr and Nd
232 isotopic ratios. Isotopic measurements were performed at the University of Arizona TIMS laboratory
233 following the procedure in Conroy et al. (2013) on soil samples. Samples were not spiked and
234 dissolved in mixtures of ultrapure Hf-HNO₃ acid. Elemental separation of dissolved samples was
235 carried out in chromatographic columns via HCl elution in a clean laboratory environment.
236 Conventional cation columns filled with AG50W-X4 resin were used for Sr and REEs separation and

237 anion columns with LN Spec resin for Nd separation following Ducea et al. (2020). Sr cuts were
238 loaded onto Ta single filaments and Nd cuts onto Re filaments. $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios (Table
239 3) were measured on a VG Sector 54 thermal ionization mass spectrometer (TIMS) fitted with
240 adjustable 1011 Ω Faraday collectors and Daly photomultipliers. NBS SRM 987 Sr standard and La
241 Jolla Nd standard were analyzed during the samples run to ensure the performance of the
242 instrument and to perform some minor correction on the final reported ratios.

243

244 **3.5 Erosional potential and aeolian dust accretion proxies**

245 The erosional potential of the three sampling sites, which are distanced along a 2km transect was
246 derived from field estimates of the vertical height of the MK and TZ rocky headwalls and their scree
247 slopes. We evaluated the plateau and scree slope energy distribution and maturity stage from the
248 dimensionless ratio between the vertical height of the scree slope (H_t) to the vertical height of the
249 headwall (H_c) following Statham (1976) (Figure 2).

250 To assess the potential contribution of distal and local aeolian dust inputs in the PM soil we used
251 the contents of quartz (wt. %). The source of quartz can be local, from the Pieria Mountains silicate
252 bedrock and from the granites to the west of Mount Olympus, or can be transported during Sahara
253 dust episodes, as evidenced from the XRD analyses of the PM snowpack samples, which are in line
254 with Sahara dust samples from the Pyrenees, the European Alps, and the Carpathian Mountains that
255 contain high amounts of quartz (e.g., Rellini et al., 2009; Rodriguez-Navarro, 2018; Marmureanu et
256 al., 2019). Herein, we cannot exclude the possibility of quartz release from the local bedrock through
257 periglacial erosion, but the amount of quartz released from local bedrock dissolution is expected to
258 be small, wt.% concentration of the insoluble residue from carbonates in Greece is less than 1%
259 (MacLeod, 1980; Kantiranis, 2001; Kirsten and Heinrich, 2022). Therefore, it is reasonable to consider
260 quartz (wt. %) as a reliable proxy of aeolian dust accretion.

261 We selected the ϵ_{Nd} ratio as a second independent proxy particularly of Sahara dust accretion
262 in the PM soil. We did not use the Sr ratio ($^{87}\text{Sr}/^{86}\text{Sr}$) as it can be impacted by the dissolution of

263 carbonate particles and replacement of Ca by Sr during pedogenetic alteration of the PM soil (e.g.,
264 Shalev et al., 2013). Sr isotopic distributions of PM soil can be further complicated by the accretion of
265 sea-salt Sr through orographic precipitation (Kurtz et al., 2001). Rain is not a significant source of Nd,
266 so the addition of rainwater and snow should not affect the Nd isotopic composition of the aeolian
267 dust, so ϵ_{Nd} is buffered against these changes (Kurtz et al., 2001). We estimated the fraction of
268 Sahara dust from the ϵ_{Nd} ratios of the PM soil, the dust deposited on the snowpack, and the local
269 bedrock following the method by Kurtz et al (2001):

270

271

$$f = \frac{(\epsilon_{Nd} \text{ PM soil} - \epsilon_{Nd} \text{ bedrock})}{(\epsilon_{Nd} \text{ Sahara dust} - \epsilon_{Nd} \text{ bedrock})} \quad (2)$$

272 As we had not obtained direct Sr and ϵ_{Nd} values from Mount Olympus bedrock, we used the value of
273 the basal sample PM 16, which is dominated by bedrock derived calcite and falls in the same value
274 range with basin-average values of terrestrial, coastal, and marine sediments deposited in the
275 Aegean Sea (Weldeab et al., 2001).

276

277 **4 RESULTS**

278 **4.1 Alpine soil formation across a hillslope energy gradient**

279 According to Statham (1976), Ht/Hc values above 0.4 characterize a mature scree slope, which is the
280 case for the TZ (0.6) but not for the MK (0.3) scree slope, and this reflects the older deglaciation age
281 of the TZ cirque (≈ 12.5 ka BP, Section 2.1). The MK scree slope is deposited behind the LIA moraine
282 (Figure 1C), so that the most recent deglaciation processes (≈ 0.6 ka BP to present) has resulted in
283 immature scree development. Conversely, the low-relief, low-erosion PM acts as a long-term
284 depocenter of slope wash and detrital (aeolian and bedrock derived through freeze-thaw action)
285 sediments. For this low-energy setting, we believe that minor colluvial contributions, cryoturbation,
286 aeolian dust accretion, fine earth translocation, and post-depositional mineral alteration are the
287 major drivers of PM soil production. The irregular boundary between the base of the PM soil and the

288 underlying regolith composed of glacial till and outwash gravels, is indicative of cryoturbation, while
289 observations of late-season soil freezing and waterlogging (Figure 4) provide permissive evidence
290 that PM soil development is disturbed by cryogenic processes. The energy gradient along the
291 contrasting environments impacts the soil color. The PM soil basal layer overlying the regolith shares
292 similar color characteristics with the MK samples and with the TZ upper sediment horizon, which
293 have grey to olive green hues (Munsell dry color 2.5– 5 Y; Table 1). Conversely, the lower clast-free
294 sediment horizon of the TZ scree shares similar Munsell dry color characteristics with the PM soil,
295 characterized by red-brown to yellow hues (7.5–10 YR, Table 1), suggesting that these soil samples
296 are more oxidized and are undergoing pedogenetic alterations.

297

298 **4.2 Grain size variation**

299 The interactions between slope processes, colluvial sediment transport, and aeolian sub additions
300 result in polymodal grain size distributions that display different shapes among MK, TZ, and PM soils.
301 Five grain size modes (M1 to M5) were mathematically derived from the application of the CFLab
302 curve-fitting algorithm. Fitting degrees were >99% and fitting residuals were <0.1%, indicating
303 excellent fits for the raw grain size distribution curves (Figure 5 A, B, C, and D). The fine-earth (clay
304 and silt) fractions resemble grain size modes M1 and M2 with respective mean grain sizes of ≈ 2 and
305 ≈ 4 μm and M3 with mean grain sizes between 10 and 30 μm . The sand fraction is composed of two
306 modal sub-populations: a fine-sand-grain size mode (M4, mean grain size ≈ 80 μm) and a coarse
307 sand-grain size mode (M5, mean grain size 440 μm) (Figure 5 E, F, G, and H; Table 1). The production
308 of coarse sand is transported to the respective interbedded sediment horizons by rockfall activity and
309 colluvial processes, or in the case of the low-sloping PM, through slope wash. The fine sand (M4)
310 subpopulation was not traced in the PM soil samples, and this can be linked to either selective
311 entrainment of M4 or to distortion of the MK and TZ grain size curves and truncation of the coarser
312 modes (Garzanti et al., 2009).

313 In addition to the distinct color variations, the contrasting slope-energy distribution between the
314 MK, TZ scree slopes and the PM depositional environments also defines their textural compositions.
315 Sediment horizons developed on the surface of the MK scree slope contain higher amounts of sand
316 ($\square 90\%$) and lower amounts of silt and clay ($\square 10\%$) compared with their TZ counterparts ($\square 75\%$ and
317 $\square 25\%$), implying that the dominance of sand in the sediment horizons of the scree slopes derives
318 from freeze-thaw and colluvial activity. The coarse-sand content (M5) of the PM soil basal layer is 6%
319 but is lower within the solum (2%–3%), suggesting either reduced periglacial activity and/or low
320 transport capacity of erosional products from the catchment through slope wash processes during
321 the PM soil formation (Table 1; Figure 2, lower graphs).

322 The grain size distribution curves of the PM soil present a significant change in shape between
323 soil depths of 14 and 16 cm, which is characterized by a 15% reduction of the clay and very-fine-silt
324 fractions (M1 and M2) and by a similar increase of silt contents (Figure 5A and B). This sharp textural
325 differentiation was not supported from field observations, where the solum appeared homogenous
326 without distinct pedogenetic horizons and without any visual evidence of an erosional layer (Figure
327 6A), but it is supported by changes in the soil color. The samples above a soil depth of 14–16 cm
328 exhibit red to brown hues (7.5 YR), whereas the samples below this layer have more yellow-red (10
329 YR) hues (Table 1). We also observed clay coatings in sparse secondary carbonates (calcretes) along
330 the lower part of the PM soil profile, which we interpret as evidence of soil mixing and downward
331 translocation of dissolved Ca and secondary calcite precipitation at the base of the soil profile. Based
332 on these observations, we partitioned the PM soil profile in two horizons: an upper Bw horizon
333 between 0 and 14 cm with red to brown hues, low clay ($\square 25\%$), and high silt ($\square 75\%$) contents, and
334 a lower illuvial Bt horizon between 14 and 32 cm with higher ($\square 40\%$) clay contents, a yellow-red hue,
335 and smaller amounts ($\square 5\%$) of sand compared with the overlying Bw horizon (Figure 6).

336

337

338 4.3 Soil and aeolian dust mineralogy

339 XRD analysis of the bulk samples reveals a mineralogy that substantially differs between the MK,
340 TZ, and PM soils and, like the soil colorization and textural variations, follows the erosional slope
341 gradient. The most dominant mineral phase in the clast-free material of the MK and TZ soils is calcite.
342 Other minerals identified include dolomite along with quartz and micas. Conversely, the bulk
343 mineralogical composition of PM soil exhibits a richer matrix of minerals that includes quartz, chlorite
344 and mixed layer clays, mica, potassium feldspars, and plagioclase (Figure 7). Calcite is dominant
345 (\square 50%) only in basal sample PM16 (Figure 6; Table 2). Quartz, clays, and mica are the most
346 dominant mineral phases in the PM soil (\square 80%) with low values in basal sample PM16, whereas
347 plagioclase, K-feldspar, and mica represent the remaining 20% (Table 2). Semi-quantitative analysis
348 of the clay mineralogy of two samples retrieved from the surface of the Bw horizon and the base of
349 the Bt horizon (samples PM1 and PM15) revealed high concentrations of smectite and kaolinite
350 (80%) and low contents of chlorite and illite. Surface sample PM1 contains 45% smectite and 35%
351 kaolinite, whereas basal sample PM15 has higher smectite (65%) and lower kaolinite (25%) contents
352 (Table 2).

353 From the comparison of the XRD spectra (Figure 7), it is obvious that the bulk mineralogy of the
354 PM soil matches that of the Sahara dust samples. Both Sahara dust samples show the presence of
355 clay minerals, quartz, mica, calcite, plagioclase, K-feldspar, and dolomite. The detected mineral
356 phases are typical of Saharan dust deposited in Europe during both dry- and wet-deposition (red
357 rains) events (Scheuven et al., 2013). Additionally, recent studies of Saharan dust wet deposition in
358 the Iberian Peninsula also indicated the presence of Fe-Ti oxides, such as goethite and hematite, and
359 of Ti oxides, such as rutile (Rodriguez-Navarro et al., 2018), but these were not depicted from our
360 XRD analyses. Despite their overall XRD spectral similarity, a pronounced difference between the
361 contemporary Sahara dust and PM soil samples is the presence of calcite and dolomite in the dust
362 samples and their near absence from the PM soil profile (Figure 7). The smooth and low intensity

363 peaks for calcite and dolomite at 29.43 and 30.7 2 θ in surface sample PM1 indicate the partial
364 removal of calcite, whereas similarly subdued peaks in basal sample PM16 denote near complete
365 decalcification of the solum (Figure 7).

366

367 **4.4 Magnetic susceptibility of PM soil**

368 The magnetic susceptibilities of the PM soil bulk samples were measured to provide insight into the
369 ferromagnetic components of the PM soil and their potential alterations. Overall, the low-frequency
370 magnetic susceptibility (χ_{lf}) is higher in the lower Bt horizon, with average values for samples PM8–
371 PM16 of $55 \times 10^{-8} \text{ m}^3 \text{ kg}^{-1}$, and lower χ_{lf} values in the Bw horizon, with average values for samples
372 PM1–PM7 of $36 \times 10^{-8} \text{ m}^3 \text{ kg}^{-1}$ (Figure 8A). Similar value ranges were measured for the high-
373 frequency magnetic susceptibility (χ_{hf}). The estimated values of frequency-dependent (χ_{FD})
374 susceptibility presenting a wide range of values ranging between 0% (sample PM13) and 14%, with
375 significantly higher values in the Bw horizon (Figure 8B). According to Dearing (1999), high χ_{FD} values
376 (>10%) are indicative of the presence of superparamagnetic Fe oxide nanoparticles (< 0.05 μm),
377 suggesting a higher amount of fine ferrimagnetic grains in the surface horizon Bw, which potentially
378 can be of detrital (aeolian and/ eroded bedrock) origin.

379 The mineral phases responsible for the magnetic enhancement of the Bt horizon were deduced
380 from high-temperature magnetic susceptibility measurements performed during a single heating–
381 cooling cycle to 700 °C (Figure 8C). We estimated the Curie temperature (T_c) of samples PM16 and
382 PM15 to examine the potential existence of superparamagnetic ultrafine particles in the base of the
383 PM soil profile, which is in contact with the regolith. The thermomagnetic analysis of sample PM16
384 failed completely, likely due to its high calcite content and absence of magnetic phases. On the other
385 hand, sample PM15 resembling the soil-regolith lower boundary, revealed a uniform χ -T behavior
386 that is indicative of the dominance of two magnetic phases (Figure 8C) – one with T_c , or
387 transformation temperature, between 260–320 °C, probably maghemite, and a second one around

388 600 °C, which is typical for oxidized magnetite (Jordanova et al., 2022). Since the nano-sized
389 pedogenic magnetite is unstable upon heating (Dunlop and Özdemir, 1997), the identified oxidized
390 magnetite suggests that weak pedogenetic production of ferromagnetic components occurs in the
391 base of PM soil profile.

392

393

394 4.5 Radiogenic isotopes

395 More information on the provenance of the PM-soil-forming material was derived from the
396 radiogenic isotope analysis of the soil samples and of the 2018 Sahara dust sample. The $^{87}\text{Sr}/^{86}\text{Sr}$
397 values of PM soil samples range from 0.71437 to 0.72071 and the ϵ_{Nd} values from -7.75 to -9.80
398 (Table 3). Overall, the PM soil $^{87}\text{Sr}/^{86}\text{Sr} - \epsilon_{\text{ND}}$ cluster together apart from sample PM16, which has the
399 lowest value of the PM soil $^{87}\text{Sr}/^{86}\text{Sr}$ ratio (Figure 9). The analyzed Sahara dust sample exhibits
400 $^{87}\text{Sr}/^{86}\text{Sr}$ value of 0.71272 that falls within the lower range of North African dust sources between
401 \square 0.71200 and 0.74000 (Erel and Torrent, 2010; Grousset and Biscaye, 2005). The Sr isotopic ratio of
402 the Sahara dust sample shows potential mixing with rainwater and local sea salt aerosols during the
403 March 2018 wet deposition event but also with other European aerosol sources, which is validated
404 by the fact that the dust plume of the March 2018 travelled over Europe before it reached Mount
405 Olympus (Figure 3A). The Sahara dust sample has an ϵ_{Nd} value of -6.80. Plotting the $^{87}\text{Sr}/^{86}\text{Sr}$ and ϵ_{Nd}
406 measurements against literature values from terrigenous, coastal, and marine sediments from the
407 Aegean Sea region (Weldeab et al., 2002) reveals an isotopic similarity between the Sahara dust and
408 of sample PM16 with these sediments (Figure 9). A reasonable interpretation of this observation
409 comes from the fact that basal sample PM16 resembles more the soil regolith and plots close to the
410 contemporary and Holocene values of Aegean Sea terrestrial, coastal and marine sediments. The
411 $^{87}\text{Sr}/^{86}\text{Sr}$ values representing the PM soil regolith show similar values with the Aegean Sea terrestrial
412 and marine sediments and likely represent a mix of Sahara dust with Mesozoic and Cenozoic bedrock
413 carbonates, which are overall characterized by low $^{87}\text{Sr}/^{86}\text{Sr}$ values of <0.70800 (Capo et al., 1998;

414 Frank et al., 2021). The two subclusters of PM soil samples have more radiogenic values compared
415 with those of the rest of the samples and clearly correspond to Bw and Bt horizons. The increasing
416 silt contents towards the surface of the PM soil profile (Figure 6, Table 1) occur with $^{87}\text{Sr}/^{86}\text{Sr}$ and ϵ_{Nd}
417 values towards more crustal values (color variation in Figure 9) that are representative of the central
418 Sahara province. Therefore, the increases in the silt fraction within the PM soil profile can be directly
419 linked to increases in Sahara dust accretion. This is further supported by range of the silt fraction
420 mean grain size between 14 and 30 μm (Table 1), which is similar to those for modern Sahara dust
421 deposits from Crete, which range 4–8 μm and 16–30 μm (Mattson and Niéhlen, 1996; Goudie and
422 Middleton, 2001). In terms of Sahara dust provenance fingerprinting, the $^{87}\text{Sr}/^{86}\text{Sr}$ and ϵ_{Nd} values of
423 the PM soil samples fall within the range (1σ) of the central North African dust source area, which
424 broadly involves the Bodele depression (PSA2; Jewell et al., 2021).

425

426 5 DISCUSSION

427 5.1 PM soil parent material

428 The mineralogical (XRD) analyses, show that calcite is the dominant mineral phase of MK and TZ
429 interbedded sandy sediment and PM basal layers (Figure 7, lower XRD diagrams TZ01 and Table 2
430 sample PM 16), which in the periglacial environment of Mount Olympus is expected to dissolve
431 slowly (e.g., Gaillardet et al., 2019) and produce an insoluble residue that comprises the PM soil
432 parent material. MacLeod (1980) analyzed the mineral composition of the insoluble residue of
433 carbonates from western Greece and defined a mineralogical suite of quartz, kaolinite, and mica
434 (illite). Kantiranis (2001) studied the carbonate rocks of northwestern Greece and found insoluble
435 residue $\square 1\text{wt.}\%$ consisting mainly of micas, quartz, hematite, chlorite, feldspars, and amphibole,
436 whereas the insoluble residue of carbonate basement rocks from Crete also resembles $\square 1\text{ wt.}\%$ of
437 the whole rock samples and is composed of a sandy loam matrix rich in quartz, plagioclase (albite),
438 and mica (illite) (Kirsten and Heinrich, 2022). Thus, the dissolution of the local carbonate parent

439 material within the interbedded sediment layers and in the basal layer of PM soil, can release very
440 small quantities of bedrock-derived impurities such as quartz, plagioclase, illite, and kaolinite that are
441 incorporated in the solum, but cannot explain the ≈ 30 cm thick PM soil mantle and ≈ 60 cm thickness
442 of the layers interbedded in the scree slopes.

443 It has also been proposed that clay in terra rossa soils can derive from isovolumetric
444 replacement of calcite to authigenic clays across a metasomatic front, but this mechanism requires
445 significant input of aeolian dust to provide essential elements such as Al, Si, Fe and K for clay
446 formation (Merino and Banerjee, 2008). Even though we did not estimate the dissolution rate of
447 Mount Olympus bedrock metacarbonates and the elemental composition of the insoluble residue,
448 we consider that the fine earth (silt and clay) contents of MK and TZ interbedded layers, which
449 average 10% and 25%, respectively, cannot be derived only by carbonate dissolution and/or by
450 isovolumetric replacement of calcite. Küfmann (2008), Krklec et al. (2022) and Ott et al. (2023)
451 propose carbonate bedrock dissolution rates between ≈ 0.23 , 0.15 and 0.4 cm/ka respectively, which
452 for the postglacial (12.5 ka BP to present) alpine soil formation on Mount Olympus imply ≈ 5 cm of
453 carbonate loss to soil formation, a value too low to explain the observed thickness of MK, TZ,
454 interbedded layers and PM soil as a result of residual clay accumulation alone. Our direct
455 observations of episodic Sahara dust deposition on the snowpack of Mount Olympus (Figure 3)
456 provide undisputable evidence of Sahara dust accretion on PM soil. The relative contribution of local
457 dust from moraines, outwash plains and from silicate bedrock formations in the vicinity of Mount
458 Olympus is estimated in the following section, but irrespective of the relative dust sources (Saharan
459 and local), the high-energy erosive regime of Mount Olympus alpine critical zone intercepts the
460 formation of extensive aeolian dust mantles, like the one found on the stable Plateau of Muses. We
461 thus suggest that the production of silt, and clay in the PM soil basal layer, partly reflects the
462 contribution of mechanically produced sandy and fine earth carbonate debris and its dissolution
463 products, which together with aeolian dust accretion, comprise the parent materials for the PM soil
464 production.

465

466 **5.2 Relative contributions of aeolian dust inputs**

467 Studies on terra rossa soils in Greece, with typical bimodal grain size distributions consisting of clay
468 and silt subpopulations with grain size ranges of 2–4 and 10–40 μm , respectively, ascribe the clay
469 fraction, which is rich in illite and kaolinite, to the limestone residue, and the silt fraction, which is
470 made up entirely of quartz, to long-range aeolian transport from variable sources (Russel and Van
471 Andel, 2003). In line with this notion, we considered the quartz wt. % content in the solum, as a
472 proxy for aeolian dust in general and not exclusively of Sahara dust. The rounded shape of quartz
473 grains observed in SEM images (Figure 11D), provides supplementary evidence for the aeolian
474 transport of quartz grains. Furthermore, we consider that the neodymium-derived mass fraction (f),
475 solely a proxy of Sahara dust accretion in the PM soil. This is supported by the high statistical
476 correlation between the silt fraction (M3) with the ϵ_{Nd} -derived f fraction ($R^2= 0.73$, $P< 0.001$) and by
477 the similarity of the grain size ranges between the silt fraction and the modern Sahara dust deposits.
478 The mass fraction (f) of Sahara-dust-derived ϵ_{Nd} was calculated based on the highest ϵ_{Nd} value of
479 sample PM 16 and Aegean Sea sediments ($\epsilon_{\text{Nd}} = -5.94$) and on the lowest value of Sahara dust PSA2
480 ($\epsilon_{\text{Nd}} = -13.81$) end members. The ϵ_{Nd} value of Aegean Sea sediments is considered conservative in
481 relation to that of Mount Olympus bedrock due to the mixing of the carbonate bedrock sediments
482 with other sources of silicate bedrock during fluvial transport.

483 The ϵ_{Nd} -based Sahara dust contributions to the PM soil varies between \square 35% and \square 50%
484 (except that of basal sample PM16) (Figure 10). Conversely, the quartz-derived aeolian dust
485 contribution ranges between \square 45% and 65%, shows a relatively small variation with depth and an
486 abrupt increase (\square 25%) from sample PM16 to PM15 (Figure 10). The basal sample PM16 exhibits the
487 lowest contributions of quartz concentration, f ratio values (Figure 10) and silt concentrations (Table
488 1, Figure 6) and is considered an outlier representing the regolith-PM soil mix, which agrees with its
489 distinct color and lowest magnetic susceptibility values. The preservation of quartz in the PM soil

490 profile and especially in the lower Bt horizon requires a mechanism of reduced Sahara dust input
491 and/or loss to weathering, with simultaneous inputs of other quartz-rich-derived dust. A pattern that
492 can explain the lower ϵ_{Nd} -based Sahara dust contributions in the Bt horizon and the near steady
493 quartz contents is a shift in atmospheric circulation patterns that resulted in less-frequent dust
494 transport episodes from north Africa along with steady aeolian quartz accretion from local quartz
495 sources. Aeolian quartz from the silicate bedrock formations of the Pieria mountains, Mount
496 Olympus granites and even from the Katerini alluvial plane (Figure 1) can be deposited on Mount
497 Olympus periglacial zone during periods of regional aridity, associated with thinning of vegetation,
498 desiccation of the Katerini alluvial plane, and immobilization of fine dust grains through convection.

499 Based on the above, we tentatively attribute the $\approx 15\%$ difference between the ϵ_{Nd} -based
500 estimates and the quartz-based estimates to accretion of quartz-rich dust from local sources during
501 the formation of the Bt horizon, considering that the contribution of bedrock derived quartz from
502 the insoluble residue is $\approx 1\%$. From the ϵ_{Nd} -based contributions, we estimate that the Sahara dust
503 accretion to PM soil is between $\approx 35\%$ to 50% , whereas local sources can potentially accrete another
504 $\approx 15\%$. Our estimated aeolian dust accretion $\approx 65\%$ is similar to the one in the North Calcareous
505 Alps, where the local contribution of dust from the periglacial zone of the North Calcareous and
506 Austrian silicate Alps is significant (Küfmann 2008), but our estimated Sahara dust contribution in the
507 PM soil is higher than its respective average contribution ($20 - 30\%$, Varga et al., 2016) in interglacial
508 soils of the Carpathian Basin. We attribute this difference to the closer proximity of Mount Olympus
509 to Sahara Desert than the Carpathian basin. Given that these values are conservative estimates, the
510 aeolian contribution may potentially be higher as, in our calculations, we have not included aeolian
511 transported micas, feldspars, and clays that are integral parts of Sahara dust samples deposited on
512 the snowpack. We thus suggest that the aeolian dust accretion comprises a minimum of $\approx 65\%$ of
513 the PM soil parent material and that carbonate bedrock erosion, and pedogenetic production of
514 detrital clays can potentially contribute another $\approx 35\%$ to the development of PM soil.

515

516 5.3 Pedogenetic alterations

517 An alternative mechanism that can explain the nearly homogeneous depth distribution of quartz
518 (Figure 10) is soil mixing by cryoturbation and subsequent translocation of fine earth particles from
519 the upper Bw to the lower Bt horizon. The mechanism of illuviation does not necessarily cancel the
520 climatic forcing of Sahara dust reduction and increase of local dust inputs during the development of
521 Bt horizon, but rather can act synergistically. For example, a cold and arid climatic phase that
522 immobilizes quartz-rich dust from the Mount Olympus and Pieria mountains piedmonts can also
523 reactivate the periglacial processes on the Mount Olympus alpine critical zone, which in turn
524 enhance scree slope aggradation, colluvial activity, intensification of freeze–thaw cycles, and
525 cryoturbation of the soils. Cryogenically induced translocation of detrital (aeolian and bedrock
526 derived) silt and clays deposited on the surface of the Bw horizon, distorts the textural composition
527 and soil properties and results in massive structures like the one we observed in the PM soil profile
528 (Figure 4).

529 Despite the absence of color difference and of distinct layers in the PM soil, the higher magnetic
530 susceptibility values of the Bt compared to the Bw horizon, can result from the enrichment of
531 ferromagnetic minerals during *in situ* weathering of translocated detrital fine earth particles through
532 pedogenesis (Maher, 2011). However, the overall low values of the frequency dependent magnetic
533 susceptibility ($\chi_{FD} < 10$), point to weak pedogenetic alteration of soil (Dearing et al., 1996), which in the
534 base of PM soil profile occurs through the oxidation of ultrafine (titano)magnetite to maghemite
535 (Section 4.4). The SEM-EDS analyses show the presence of ultrafine Fe-Ti grains throughout the
536 solum (Figure 11B) apart from basal sample PM16 (Figure 11A), which is representative of the PM
537 soil regolith. This is further supported from the EDS chemical composition of the calcite grains in
538 basal sample PM 16 that have TiO₂ weight % concentration <1%. On the other hand, magnetite is
539 found attached to clay minerals of the Bodéle depression surface sediments (Moskowitz et al., 2016),
540 which is the major source of Sahara dust in PM soil (Figure 10). Also, magnetic susceptibility

541 measurements of Sahara dust modern deposits in SE Bulgaria show a low frequency magnetic
542 susceptibility value of $\chi_{lf}=97 \times 10^{-8} \text{ m}^3 \text{ kg}^{-1}$ (Jordanova et al., 2013), which is close to the values of Bw
543 horizon ($\chi_{lf}=86 \times 10^{-8} \text{ m}^3 \text{ kg}^{-1}$) and Sahara dust episodes are known to transport Fe-Ti oxides in the
544 Mediterranean region (Rodriguez-Navarro et al., 2018). We can therefore ascribe the observed Fe-Ti
545 oxides and (titano)magnetite an aeolian origin, from either the local igneous silicate outcrops, or the
546 Sahara Desert. Collectively these observations imply that the magnetic enhancement of Bt compared
547 to Bw horizon can result from a combination of fine earth illuviation of aeolian transported ultrafine
548 magnetic particles from the Bw horizon to the Bt horizon through cryoturbation and subsequent
549 weak pedogenetic modifications that result to the oxidation of magnetic minerals like
550 (titano)magnetite and can also explain the reddish to yellow color hues.

551

552 **5.4 Mineral weathering**

553 In addition to the weak pedogenesis of ferromagnetic minerals in the base of the PM soil profile, we
554 assessed the mineral weathering potential of non-magnetic minerals through the clay mineralogy
555 composition of basal and topsoil samples PM15 and PM1. Both samples show the dominance of
556 smectite with lesser contributions by kaolinite, chlorite, and illite (Table 2). High amounts of smectite
557 in alpine soils result from the alteration of detrital chlorite and micas deposited on glacier surfaces
558 and are found in proglacial fields in the European Alps and Rocky Mountains (Egli et al., 2003; Egli et
559 al., 2011; Munroe et al., 2015), so that the 20% difference in smectite concentration (Table 2)
560 between the basal and topsoil layers of PM soil can be partly related to enhanced mineral chemical
561 weathering in the base of the solum. Similarly, kaolinite observed in the XRD profiles of the MK, TZ
562 and Sahara dust samples (Figure 7), can be released from the dissolution of bedrock carbonates, as
563 are the cases for western Greece (Macleod, 1980) and Crete (Kirsten and Heinrich, 2022), but can
564 also form through the alteration of other detrital minerals, such as plagioclase (albite), a process that
565 is common in glacial and periglacial environments (Anderson, 2000). Finally, high smectite and

566 kaolinite contents can also be transported during Saharan dust transport episodes (e.g., Scheuven et
567 al., 2013), but specifically they are representative of the western Sahara dust provenance (PSA 1, Figure
568 10; Rodriguez-Navarro et al., 2018). However, smectite and kaolinite are also found in modern
569 Sahara dust samples deposited in Athens, Greece (Remoundaki et al., 2011). Therefore, we consider
570 that the high (>80%) concentration of smectite and kaolinite in the PM soil clay (< 2 μm) fraction
571 reflects the balance between direct aeolian deposition and *in-situ* weathering of detrital (aeolian
572 and/or bedrock derived) micas and plagioclase, but the respective contributions of aeolian-
573 transported versus that of bedrock-derived clay minerals subjected to post-depositional mineral
574 alterations cannot be defined from the existing data.

575

576 **5.5 Relative timing of PM soil development**

577 Direct observations suggest that cryoturbation is a fundamental pedogenetic process in the
578 development of PM soil and continues today along with the ongoing accretion of the surficial aeolian
579 silt horizon Bw. The occurrence of seasonal soil freezing and lack of vegetation in the PM polygon
580 centers provide evidence that cryoturbation is active, destroying soil horizonation and obscuring
581 pedogenetic and chemical weathering signals. However, magnetic, and mineralogical data indicate
582 the occurrence of weathered Fe-(Ti) oxides such as (titano)maghemite, and the dominance of
583 smectite and kaolinite in the soil basal and topsoil layers, which enable us to conclude that mineral
584 alteration, and pedogenetic modifications of deposited aeolian dust and local erosional products are
585 ongoing processes within the PM soil profile, occurring in tandem with cryoturbation.

586 In the absence of absolute datings that can constrain temporally the processes driving the
587 production of PM soil, we hypothesized on its age based on the conclusions drawn from the
588 contributions of aeolian dust, and the impacts of cryoturbation. We tentatively ascribe the
589 deposition of the base colluvial layer and/or the *in-situ* fragmentation of the regolith's till boulders to
590 the most recent period of glacial activity on Mount Olympus. Based on the glacial record of the MK

591 and TZ cirques, the best candidates of periglacial activity that have likely resulted in the deposition of
592 outwash sand and gravels postdate the moraine stabilization phases at ≈ 12.5 , 2.5, and 0.6 ka BP.

593 However, there is a 10-ka time span between the Holocene–Pleistocene boundary and the
594 late-Holocene glacial expansions on Mount Olympus. Accepting that PM soil formation began after
595 the moraine stabilization phase at ≈ 12.5 ka BP that was common to MK and TZ cirques, its
596 production rate would be $\approx 3 \times 10^{-5}$ m yr⁻¹ assuming that soil erosion in the low-lying PM has been
597 minimal. This rate is considerably lower than respective soil production rates of Alpine and
598 Mediterranean soils formed over the last 10 ka (Egli et al., 2018; Figure 8). In contrast, by considering
599 a late-Holocene age and that the PM soil development postdates the ≈ 2.5 ka BP moraine
600 stabilization phase, the soil production rate is $\approx 1 \times 10^{-4}$ m yr⁻¹ an estimate that is in better
601 agreement with the soil production rates presented by Egli et al. (2018) for both Alpine and
602 Mediterranean soils. Furthermore, a late-Holocene development of PM soil broadly agrees with soil
603 development patterns in diverse geomorphological environments in Crete (Kirsten and Heinrich,
604 2022). If this scenario is correct, then we can further hypothesize that development of the Bt horizon
605 could have lasted between ≈ 2.5 and 1.0 ka BP, before a recorded phase of intense Sahara dust
606 accretion in Mediterranean that resulted from the combined action of an orbitally induced decrease
607 in solar insolation and of increased aridity over North Africa (Sabatier et al., 2020). This shift could
608 potentially explain the sharp textural boundary between the Bt and Bw horizons and the increasing
609 Sahara dust accretion on the upper Bw horizon. The hypothesized development of the Bw horizon
610 over the past 1 ka could have been disturbed by cryoturbation during the LIA (≈ 0.6 ka BP) glacial
611 expansion in the MK and that continues until today. Ongoing work on Mount Olympus alpine critical
612 zone involves efforts to accurately date the MK and TZ scree interbedded layers and the PM soil
613 profile through Optically Stimulated Luminescence dating that is aided by the high concentrations of
614 quartz in the fine earth fraction, as well as additional geochemical analysis and estimates of the local
615 carbonate bedrock dissolution rates and its residual geochemical composition, in an overall attempt

616 to provide a new continuous record of postglacial alpine landscape evolution in the Mediterranean
617 periglacial zone.

618

619 **6 CONCLUSIONS**

620 In this study, we investigated the local processes that lead to the development of alpine soils on a
621 stable landform on Mount Olympus, considering its regional setting representative of Mediterranean
622 carbonate mountains that became gradually ice-free during the Pleistocene–Holocene transition but
623 that have also been affected by late-Holocene climatic shifts towards glacial and periglacial
624 conditions (Oliva et al., 2018). We discussed the relative contributions of erosion, aeolian dust
625 accretion, and post-depositional pedogenesis and mineral alteration by comparing colluvial sediment
626 layers interbedded in scree slopes with a soil B horizon developed on a regolith composed by slope
627 outwash deposits and fragmented till boulders along a 2km hillslope energy gradient with a
628 northeasterly orientation, which is the main direction of glacial cirque development on Mount
629 Olympus.

630 Overall, our results suggest that soils developed in stable landforms like the PM show signs of
631 weak pedogenesis and contain higher amounts of aeolian dust than locally eroded and chemically
632 weathered products. Aeolian dust from local and Saharan sources is accreted in alpine soils formed in
633 periglacial hummocky polygons of the PM and comprises \square 30%–65% of the soil mass weight. This
634 interpretation matches those of several other studies on aeolian dust accretion in alpine soils (e.g.,
635 Gild et al., 2018; Kaüfmann, 2008; Munroe et al., 2015; Yang et al., 2016; Kirsten and Heinrich, 2022)
636 and suggests that aeolian dust is the primary parent soil material on Mount Olympus. The major
637 source of Sahara dust deposited on Mount Olympus is the Bodélé depression, which agrees with
638 observations of accreted dust in Crete (Pye, 1992).

639 In the low-erosional environment of the PM, mineral alteration and weak pedogenetic
640 modifications occur throughout the solum, but their signal is blurred by soil mixing due to ongoing

641 cryoturbation. A sharp textural boundary not visible in the field separates an upper weathered soil
642 Bw horizon from the lower Bt horizon, which is magnetically enhanced and enriched in smectite and
643 kaolinite. Radiogenic isotope systematics, mineralogy, and magnetic susceptibility value range
644 classify the Bw horizon as an aeolian silt layer that was likely formed during a late-Holocene shift of
645 regional atmospheric circulation that resulted in increased Sahara dust accretion in alpine
646 Mediterranean landscapes.

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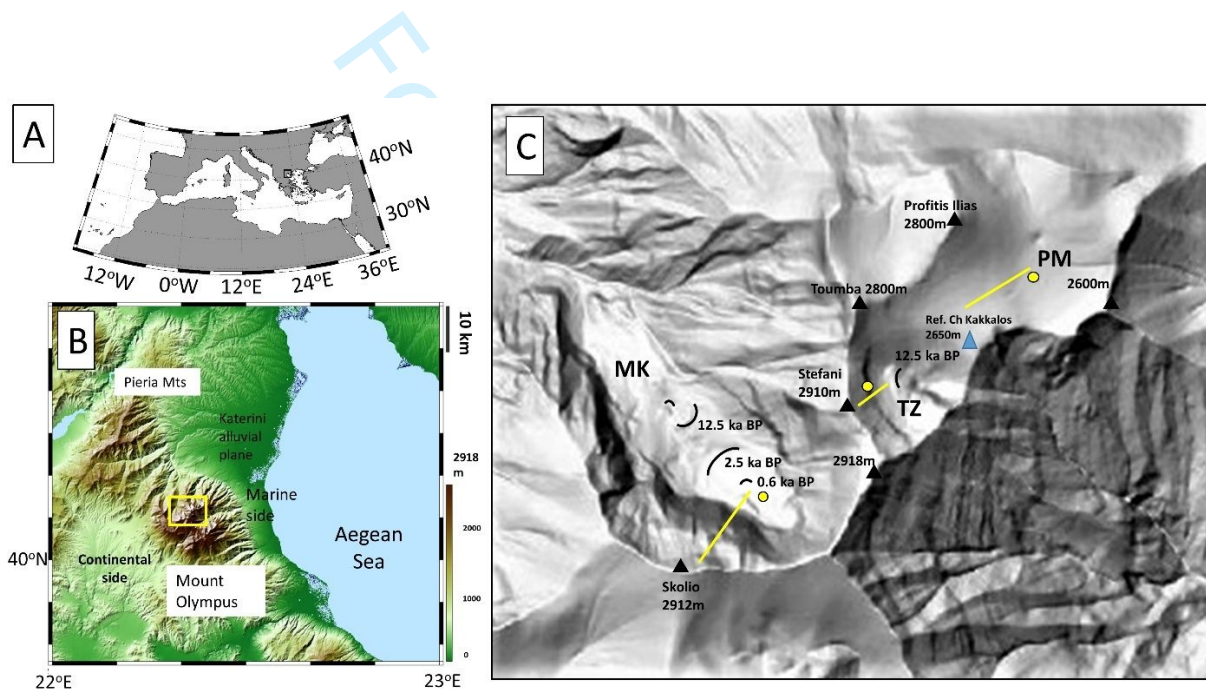
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874 **FIGURES**

875 **Figure 1.** (A) General setting of the study area within the Mediterranean basin. (B) Mount Olympus
 876 alpine domain that is considered in the study (yellow box), with the two respective piedmonts on the
 877 marine and continental sides, the adjacent Pieria Mountains and Katerini alluvial plane (SRTM 90
 878 DEM Model). (C) The highest cirques and plateau of Mount Olympus with the respective locations of
 879 dated moraines (black curved lines from Styllas et al., 2018), the sampling locations considered in this
 880 study (yellow circles) and the geomorphological transects described in Figure 2 (yellow lines). MK:
 881 Megala Kazania cirque, TZ: Throne of Zeus cirque, PM: Plateau of Muses.

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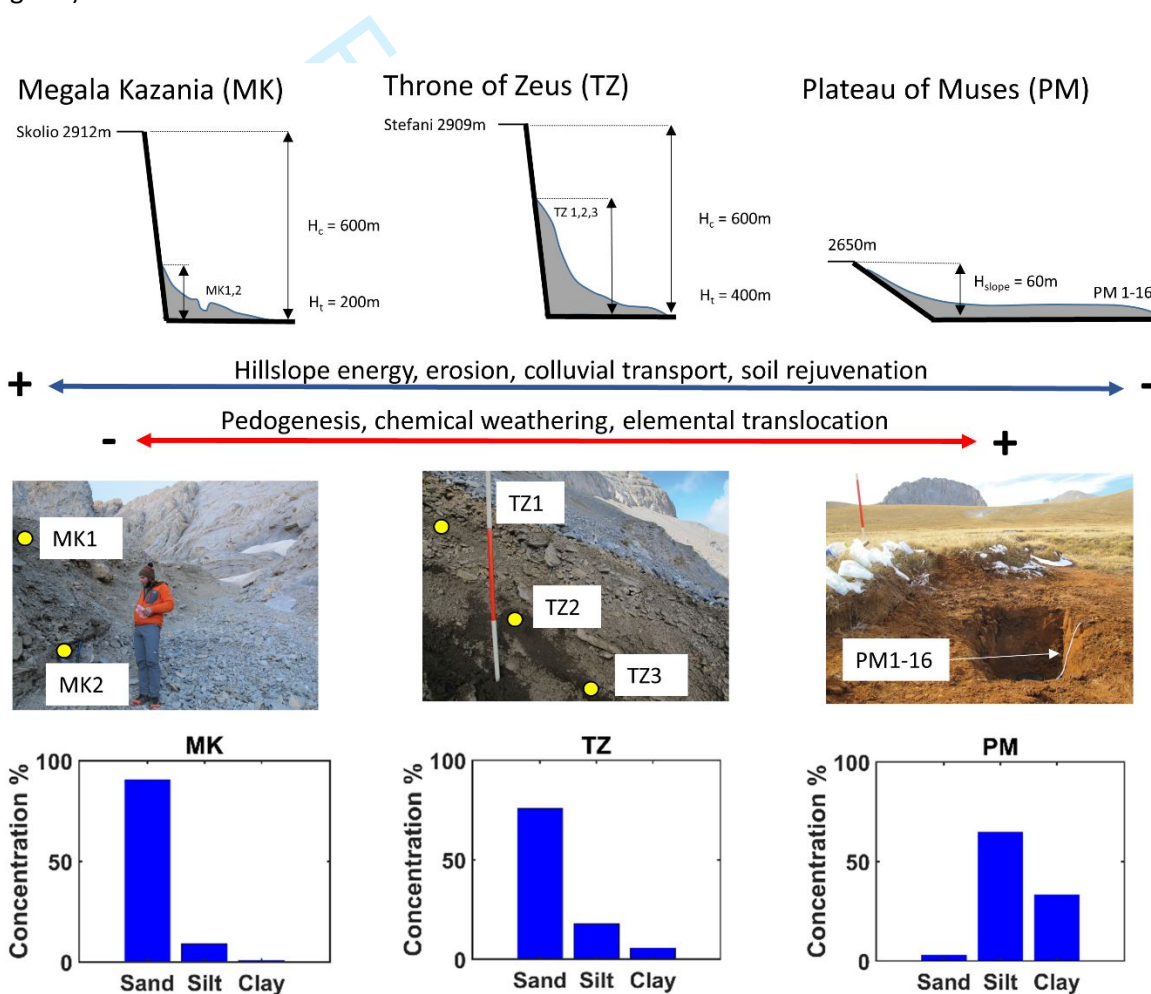


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886 **Figure 2.** Conceptual diagram of the study, with the sampling sites and their morphological profiles,
 887 shown in Figure 1c as yellow lines, and with their respective textural characteristics that resulted
 888 from the grain size analysis. The respective heights of the rock cliffs (H_c) and talus slopes (H_t) are
 889 shown. The soil samples from the stratified scree clast free horizons in MK cirque are located behind
 890 the Little Ice Age moraine (left upper panel, photo, and diagram). The stratified scree slope under the
 891 rock wall of Stefani (2910m) in the TZ cirque, with the respective locations of the clast free soil
 892 samples (center panel, photo, and diagram) and the soil profile in the PM (right panel, photo, and
 893 diagram).



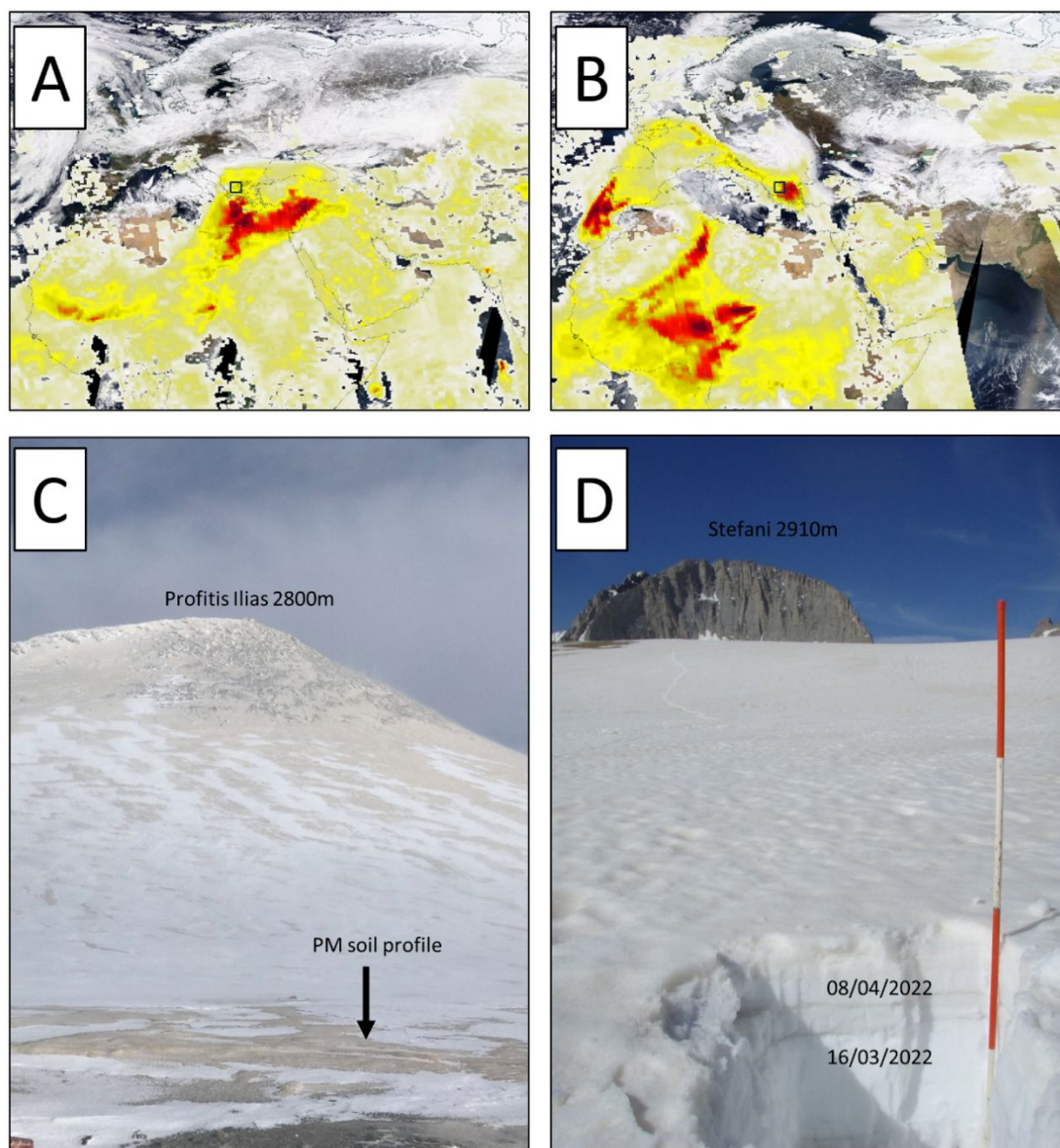
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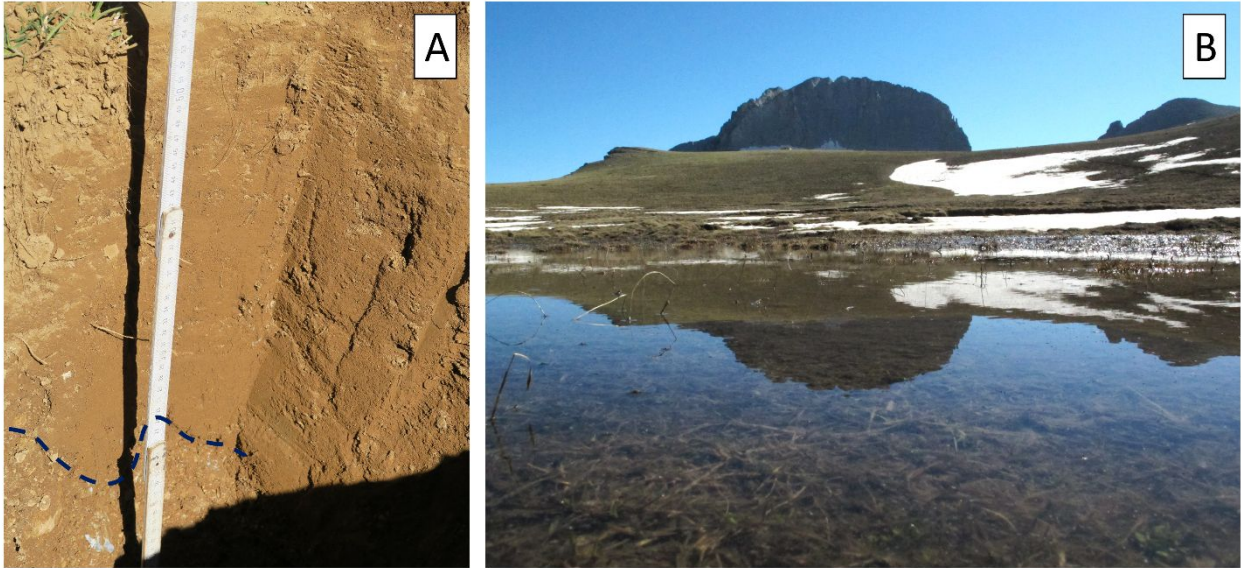
897 **Figure 3.** Synoptic maps and direct observations of two Sahara dust episodes on Mount Olympus
 898 alpine critical zone (black rectangle). (A). Aerosol Optical Depth (AOD) during March 22, 2018, and (B)
 899 and March 16, 2022, showing the trajectory of dust plume from Sahara Desert, and their impacts on
 900 the snowpack of the Plateau of Muses (C and D). The PM soil profile was excavated under the black
 901 arrow (C), whereas the snow pit (D) with two successive Sahara dust transport episodes in the spring
 902 of 2022, has also been excavated on top of the PM soil profile excavated pit. The NASA SUOMI/NNP
 903 Aerosol Optical Depth composition product was downloaded from the NASA EOSDIS Worldview
 904 platform (worldview.earthdata.nasa.gov).

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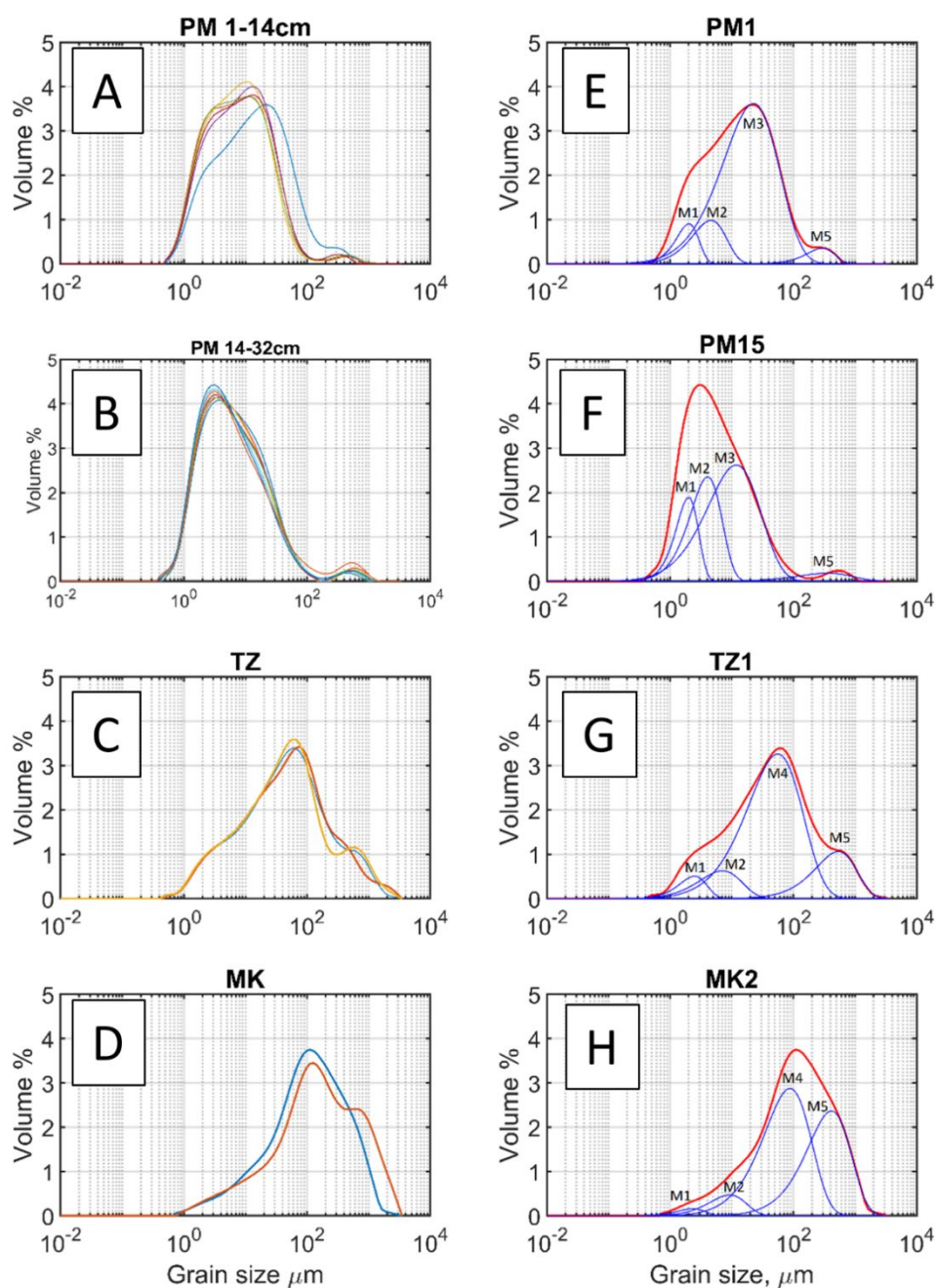
907 **Figure 4.** Evidence of soil disturbance on the Plateau of Muses under past and present-day climatic
908 conditions. (A) An irregular gravel layer (blue dashed line) between colluvial gravel and the overlying
909 soil resulting from cryoturbation. (B) Early summer season ongoing freeze of the soil surface layer
910 and subsequent waterlogging (photo taken on June 2012).



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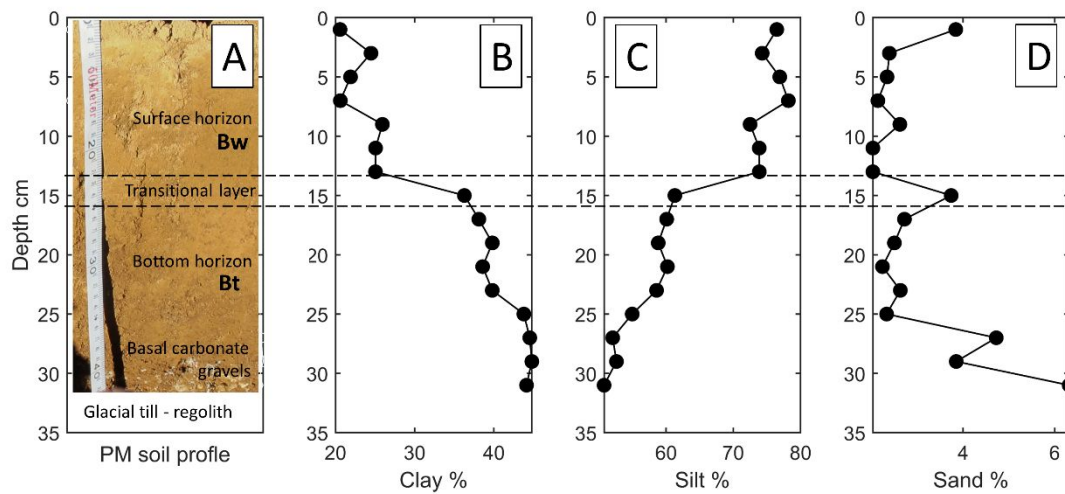
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913 **Figure 5.** Cumulative grain size distributions of the soil samples from MK, TZ and PM Bw and Bt
 914 horizons (A, B, C and D). Surface sample PM1 (A, blue line) shows a distinct grain size distribution
 915 from the PM soil upper layer. Subplots E, F, G and H: Results of the CFLab fitting algorithm with the
 916 respective grain size distributions (GSD) and extracted grain size modes M1 to M5 (blue curves) of
 917 the soil samples PM1, PM15, TZ1 and MK2, represented as distinct sub-populations.



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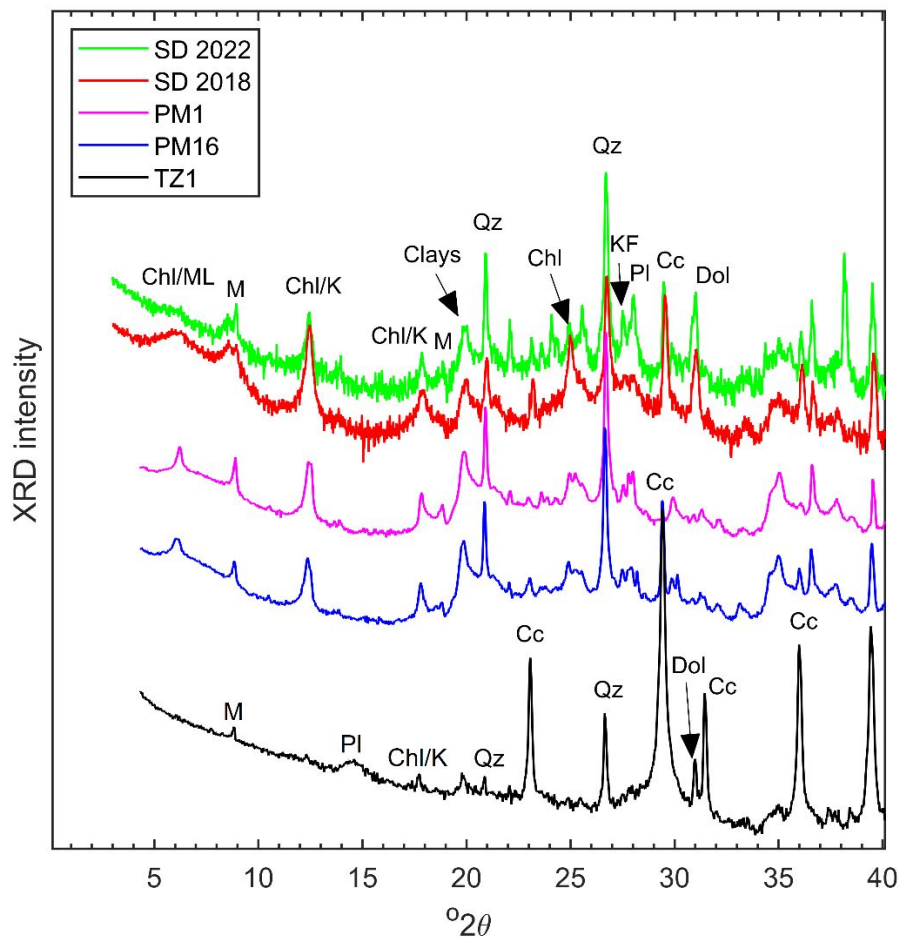
919 **Figure 6.** The PM soil profile along with the depth variations main textural classes. A transition layer
 920 between 14 and 16 cm of depth marks a substantial decrease in clay and increase in silt contents and
 921 a change in the dry soil color. For clarity reasons, it is noted that the direct depth measurement of
 922 the PM soil begins at 10 cm along the measurement tape, explaining the discrepancy between the
 923 actual and the illustrated depth. The photo was taken one day after the profile excavation, when the
 924 upper part had partly dried out.



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928 **Figure 7.** X-ray diffraction patterns of soil samples PM1, PM15, TZ1 and MK2. Soils within the
 929 hillslope high energy scree deposits are composed primarily of calcite. In contrast PM soil samples
 930 contain quartz, clays, feldspars, and mica. (M: mica, Chl/M: Chlorite and mixed layer clays, likely
 931 smectite, Chl/K: chlorite and likely kaolinite, K: K feldspar, Qz: quartz, Pl: plagioclase feldspars, Cc:
 932 calcite, Dol: Dolomite).



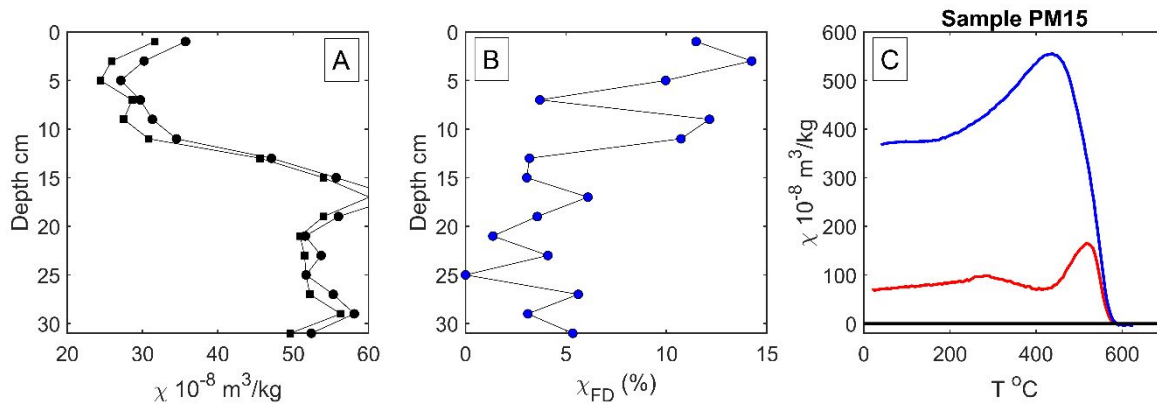
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937 **Figure 8.** Depth variations of low and high frequency (A) and frequency dependent magnetic
 938 susceptibility (B) and (C) thermomagnetic analysis results of sample PM15 (red heating curve, blue
 939 cooling curve).



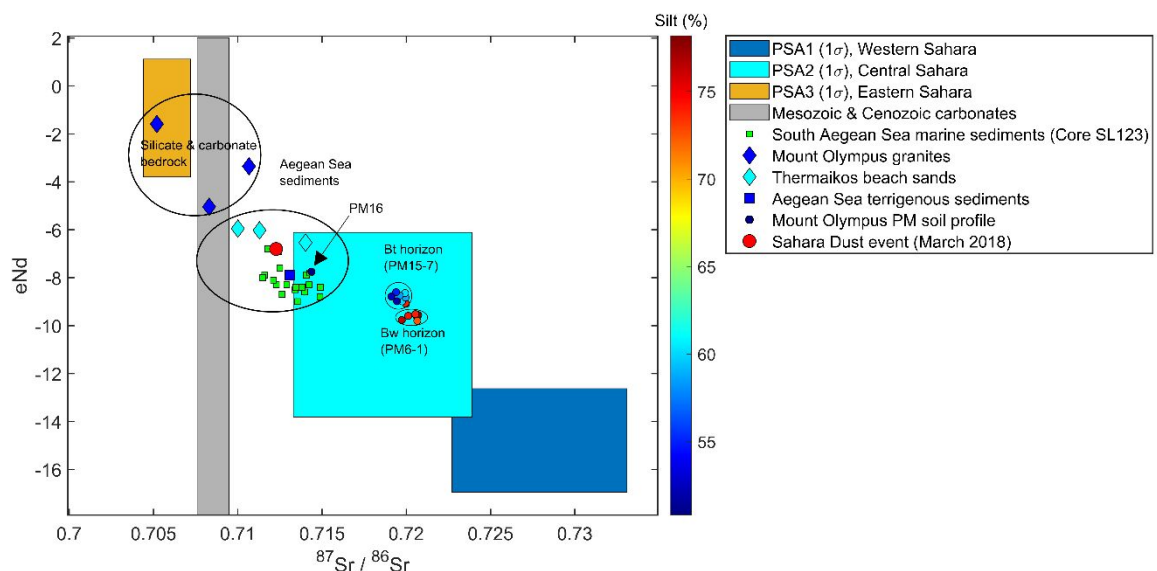
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943 **Figure 9.** Plot of $^{87}\text{Sr}/^{86}\text{Sr}$ against ϵ_{Nd} values of the PM soil samples with respective values of Mesozoic
 944 and Cretaceous carbonates (Franck et al., 2021), Mount Olympus granites (located on the continental
 945 west sides of the massif (Šarić et al., 2009, Castorina et al., 2020), Aegean Sea terrigenous and coastal
 946 sediments (Weldeab et al., 2002) and South Aegean marine sediments from Core SL 123 (Ehrmann et
 947 al., 2007), along with the three main North African dust source areas (PSA, Jewell et al., 2021). The
 948 isotopic enrichment trend for the PM soil samples towards crustal more radiogenic values occurs
 949 with a 25% increase in silt contents (colorbar) from the base to the surface of PM soil profile,
 950 suggesting the influence of external aeolian dust.

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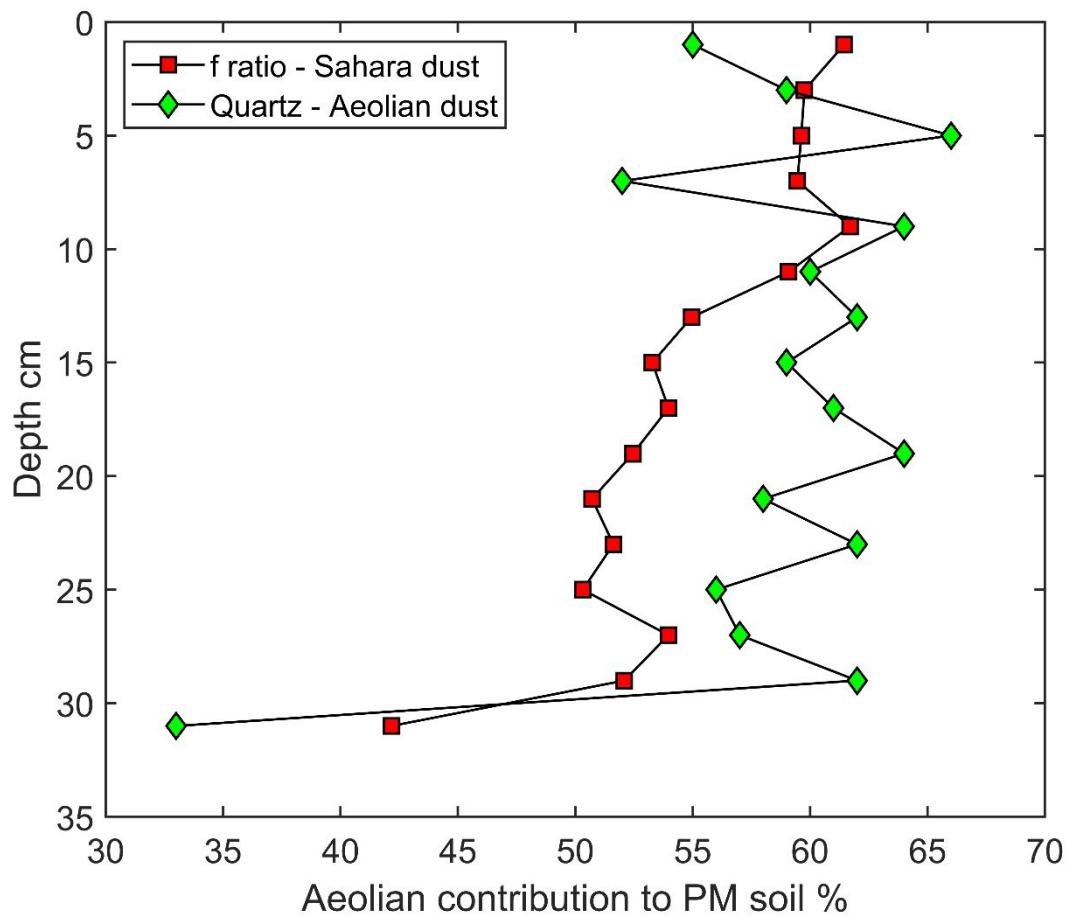
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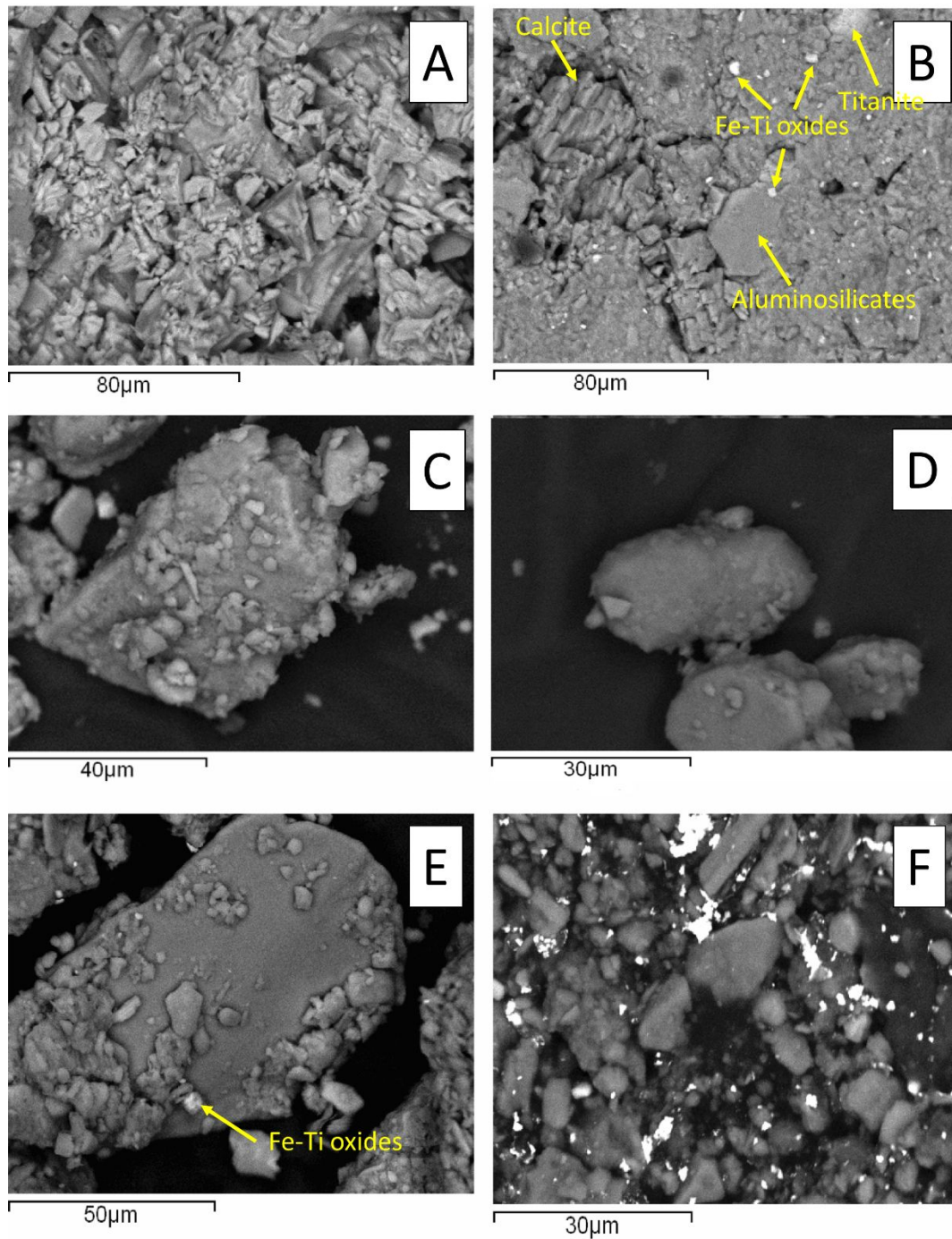
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958 **FIGURE 10.** Estimates of the relative contributions of aeolian dust accretion to PM soil as calculated
 959 by mineralogical, and isotopic proxies.



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968 **FIGURE 11.** SEM backscatter images from selected samples of PM loess profile. (A) Calcite grains
 969 from basal sample PM 16. (B) Mixed phase of aluminosilicates with calcite, titanomagnetite and
 970 titanite from basal sample PM 16. (C) K-feldspar. (D) Quartz grain with rounded edges as a result of
 971 long-range aeolian transport. (E) Surface sample PM1 aggregate of aluminosilicates and Fe-Ti oxides.
 972 (F) Quartz grains of variable shapes and grain sizes from sample PM3 along with Fe-Ti oxides.



973

974 **Table 1.** Physical characteristics of the soil samples retrieved from the interbedded colluvial soils of
 975 the Megala Kazania (MK) and Throne of Zeus (TZ) scree slopes and from the alpine soil formed on the
 976 Plateau of Muses (PM).

Sample id	Depth (cm)	Munsell color (dry)	Clay (%) ($<2\mu\text{m}$)	Fine Silt (%) ($3.5\text{-}5\mu\text{m}$)	Silt (%) ($14\text{-}30\mu\text{m}$)	Fine sand (%) ($65\text{-}110\mu\text{m}$)	Coarse sand (%) ($300\text{-}800\mu\text{m}$)
MK1	30	5Y 6/1	1.7	6.6	0.0	39.8	51.9
MK2	250	5Y 6/1	0.0	11.4	0.0	47.5	41.1
TZ1	50	2.5Y 5/4	5.3	10.0	0.0	67.9	16.8
TZ2	150	10YR 3/4	6.0	25.2	0.0	43.5	25.3
TZ3	120	10YR 3/4	5.8	18.2	0.0	56.9	19.1
PM1	0-2	7.5YR 3/6	7.8	12.7	76.5	0.0	3.0
PM2	2-4	7.5YR 3/6	9.8	14.7	74.5	0.0	1.0
PM3	4-6	7.5YR 3/6	8.5	13.4	76.9	0.0	1.2
PM4	6-8	7.5YR 3/6	8.0	12.6	78.2	0.0	1.2
PM5	8-10	7.5YR 3/6	9.8	16.1	72.5	0.0	1.6
PM6	10-12	7.5YR 2/4	9.3	15.7	73.8	0.0	1.2
PM7	12-14	7.5YR 3/6	9.3	15.7	73.8	0.0	1.2
PM8	14-16	10YR 3/4	13.0	23.3	61.3	0.0	2.4
PM9	16-18	10YR 3/4	13.8	24.2	60.1	0.0	1.9
PM10	18-20	10YR 2/4	14.5	25.2	58.8	0.0	1.5
PM11	20-22	10YR 3/4	14.0	24.5	60.2	0.0	1.3
PM12	22-24	10YR 3/6	14.6	25.1	58.6	0.0	1.7
PM13	24-26	10YR 3/6	16.3	27.5	55.0	0.0	1.2

PM14	26-28	10YR 3/4	16.0	28.5	52.1	0.0	3.4
PM15	28-30	10YR 4/6	16.2	28.6	52.6	0.0	2.6
PM16	30-32	2.5YR 5/6	15.6	28.5	50.8	0.0	5.1

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979 **Table 2.** Weight percent (wt %) mineralogical semi quantitative composition of the PM soil, along
 980 with the clay mineralogy of surface and base samples PM1 and PM15.

Sample id	Qtz	Chl_CC	Plag	KF	Mica	Amph	Cc	Clay mineralogy (<2µm) Sm / Kaol / Chl / Ill
PM1	55	21	14	3	7	0	0	Smectite: 45%, Kaolinite: 35%, Chlorite: 10%, Illite:10%
PM2	59	26	6	1	8	0	0	
PM3	66	21	4	2	7	0	0	
PM4	52	31	7	3	7	0	0	
PM5	64	22	7	0	7	0	0	
PM6	60	24	7	2	7	0	0	
PM7	62	20	4	6	8	0	0	
PM8	59	24	7	3	7	0	0	
PM9	61	24	7	3	5	0	0	
PM10	64	22	8	1	5	0	0	
PM11	58	28	4	3	7	0	0	
PM12	62	22	7	2	7	0	0	
PM13	56	25	5	1	7	6	0	
PM14	57	28	2	7	6	0	0	
PM15	62	24	6	4	4	0	0	Smectite: 65%, Kaolinite: 25%, Chlorite: 5%, Illite:5%
PM16	33	12	1	2	4	0	48	

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982

983 **Table 3.** Radiogenic isotope results for the PM soil profile and the 2018 Sahara dust (SD) samples.

984 ϵ ND values were calculated

Sample id	$^{87}\text{Sr}/^{86}\text{Sr}$		$^{143}\text{Nd}/^{144}\text{Nd}$		ϵ_{ND}
	$^{87}\text{Sr}/^{86}\text{Sr}$	std err (%)	$^{143}\text{Nd}/^{144}\text{Nd}$	std. err. (%)	
NBS 987 Sr standard	0.7102500	0.0008	-	-	
La Jolla Nd standard	-	-	0.5118500	0.0006	
PM1	0.7197322	0.0008	0.5121292	0.0004	-9.77
PM2	0.7201146	0.0026	0.5121383	0.0006	-9.60
PM3	0.7205816	0.0009	0.5121390	0.0006	-9.58
PM4	0.7207105	0.0010	0.5121399	0.0007	-9.56
PM5	0.7206530	0.0029	0.5121278	0.0009	-9.80
PM6	0.7205393	0.0007	0.5121419	0.0005	-9.52
PM7	0.7199828	0.0008	0.5121641	0.0005	-9.09
PM8	0.7196993	0.0009	0.5121731	0.0006	-8.91
PM9	0.7197594	0.0008	0.5121694	0.0005	-8.98
PM10	0.7199571	0.0007	0.5121775	0.0004	-8.82
PM11	0.7199244	0.0009	0.5121869	0.0006	-8.64
PM12	0.7196162	0.0008	0.5121820	0.0006	-8.74
PM13	0.7193930	0.0008	0.5121890	0.0006	-8.60
PM14	0.7194477	0.0009	0.5121694	0.0005	-8.98
PM15	0.7191162	0.0011	0.5121795	0.0004	-8.79
PM16	0.7143748	0.0016	0.5122328	0.0005	-7.75
SD 2018	0.7122721	0.0009	0.5122813	0.0006	-6.80

985

1 **Aeolian dust accretion outpaces erosion in the formation of Mediterranean**
2 **alpine soils. New evidence from the periglacial zone of Mount Olympus,**
3 **Greece**

4
5 **Abstract**

6 Soil formation in Mediterranean periglacial landscapes remains poorly understood as the interplay
7 between erosion and aeolian dust accretion in providing parent materials, and mineral weathering
8 and pedogenesis, as dominant post depositional processes, depends on a variety of local and
9 regional factors. Herein, we investigate the balance between erosion and aeolian dust accretion in
10 the formation of an alpine soil profile along a 2km erosional gradient of decreasing power in the
11 northeastern Mediterranean alpine hinterland and specifically in the periglacial zone of Mount
12 Olympus in Greece. We applied a wide range of analytical methods to 23 samples, from a soil profile
13 developed in a glaciokarstic plateau, from sediment horizons interbedded in postglacial scree slopes
14 and from modern Sahara dust samples deposited on the snowpack. Clast free horizons developed on
15 scree slopes exhibit high concentrations of calcite rich sand and are representative local erosion
16 products. The alpine soil B horizon developed on a glaciokarstic plateau contains high amounts of
17 fine earth and is rich in quartz, mica, plagioclase, clays, and Fe-Ti oxides. Based on its physical and
18 textural characteristics the soil profile is partitioned in a surficial weathered Bw horizon and a lower
19 illuvial Bt horizon that overlies the local regolith composed of fragmented glacial till and slope wash
20 sand and gravels. Radiogenic isotope systematics, grain size and mineralogical analysis show that the
21 contribution of Sahara and locally sourced dust to the development of the soil B horizon ranges
22 between 50 and 65%. Cryoturbation results in fine earth translocation from Bw to the Bt horizon,
23 whereas weak pedogenetic modifications of detrital (aeolian and bedrock-derived) minerals result in
24 magnetic mineral weathering and secondary clay (smectite and kaolinite) formation. Our findings
25 reveal that, in addition to the low dissolution potential of the local regolith, aeolian dust accretion is
26 the dominant process in providing alpine soil parent material and that cryoturbation, weak

Commented [A1]: The title has changed as the main question raised by this work is to examine the relative balance of aeolian dust accretion and local erosion in the periglacial zone of Mount Olympus in Greece, following the comments of Reviewr 1

27 pedogenesis, and mineral alteration occur within the Mediterranean periglacial zone of Mount
28 Olympus.

29

30 **Keywords:** alpine soil; erosion; aeolian dust accretion; mineral weathering; Mediterranean periglacial
31 zone, Mount Olympus

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34 1. INTRODUCTION

35 Global glacier retreat and the melting of permafrost and ground ice have altered the dynamics of the
36 alpine critical zone by enhancing erosion and by disturbing the production of mountain soils
37 (Haeberli et al., 2006, Egli et al., 2014). During periods of glacial retreat and paraglacial adjustment,
38 alpine soils develop from parent materials sourced through a combination of frost shattering,
39 colluvial activity, and hillslope outwash (Egli and Poulleard, 2016). An equally important factor that
40 affects the formation and evolution of alpine soils is the accretion of local and long-range
41 transported aeolian dust (Muhs and Benedict, 2006; Küfmann 2008; Lawrence et al., 2013; Drewnik
42 et al., 2014; Yang et al., 2016; Gild et al., 2018; Munroe et al., 2019). Thus, the contributions of
43 physical erosion and aeolian dust accretion are fundamental sources of alpine soil parent material
44 and largely define their textural, mineralogical, and geochemical characteristics.

45 The postglacial adjustment of alpine valleys is inherently linked to high rates of erosion, with
46 frequent rockfalls, debris flows, rock avalanches, and high rates of sediment production especially
47 below steep rockwalls. In such dynamic environments, alpine soil mantles formed on the surface of
48 slope deposits are patchy, often truncated and constantly rejuvenated from rockfall material,
49 whereas the evolution of these soils alternates between progressive and regressive phases (Egli et
50 al., 2018). Similar soil mantles developed on sandy layers deposited on the surface of stratified scree
51 slopes are generally indicative of quiescent periods of slope processes and are thus concise indicators
52 of optimum climatic conditions and alpine landscape stability (Sanders et al., 2010). When the
53 regional climate shifts to a colder regime, intense freeze–thaw activity and frost cracking enhance
54 rockfall activity and result in the erosion and gradual burial of these incipient soil mantles. As
55 hillslope processes and scree slope aggradation diminish away from the alpine steep rockwalls, the
56 development of alpine soils on distal moraines, outwash plains, and glacially scoured plateaus can be
57 considered continuous (*sensu lato*). In these depositional environments, low erosional rates provide
58 ample time for pedogenetic processes such as chemical weathering, mineral alteration, elemental
59 translocation, and illuviation to occur, whereas other physical processes, such as cryoturbation

Commented [A2]: Removed 'discontinuous' soil-sediment accumulations (hereafter 'soils') and replaced with 'soils'

Commented [A3]: Illuviation is included to pedogenetic processes

60 disturb the soil profiles. Alpine soils are an important component of high mountain ecosystems, so a
61 better understanding of the processes that drive their formation in climatically sensitive regions, like
62 the Mediterranean, is required.

63 Most soils formed in the Mediterranean basin display a distinguishable red color (terra rossa)
64 that derives from high concentrations of ultra-fine pedogenic iron oxides, mainly hematite (Yaalon,
65 1997; Durn et al., 1999). Terra rossa soils receive significant aeolian dust additions from Sahara and
66 Sahel regions (Yaalon, 1997; Durn, 2003; Stuut et al., 2009). In the Mediterranean alpine hinterland,
67 thin drapes of Sahara-dust-rich soils are found on plateaus, glacial moraines, and outwash plains
68 (e.g., Rellini et al., 2009), whereas aeolian dust accretion in terra rossa soils can also originate from a
69 wide range of alluvial deposits, such as sand dunes, desiccated alluvial planes, and Quaternary loess
70 (Amit et al., 2020; Lehmkuhl et al., 2020). Most of the Mediterranean mountains are built up by
71 carbonate rocks, hence the aeolian input to alpine soil formation occurs in parallel with colluvial
72 deposition of carbonate erosion and dissolution products that form a characteristic insoluble residue
73 incorporated in the soil sequences (Durn 2003; Varga et al., 2016; Kirsten and Heinrich, 2022).

74 In the present study, we investigate the major processes that drive the postglacial formation of
75 Mediterranean alpine soils in the periglacial landscapes of Mount Olympus, Greece. We follow a
76 combined sedimentological, mineralogical, and isotopic approach, and present a detailed
77 characterization of distinct alpine sediment and soil horizons developed across a geomorphological
78 gradient of decreasing erosive power. Discrete sediment samples from intact sandy layers
79 interbedded in postglacial stratified scree slope deposits that represent *in situ* erosional products of
80 the periglacial zone of Mount Olympus, are compared with samples from a soil profile developed in a
81 glaciokarstic plateau, with a goal to assess the relative contributions of aeolian dust accretion to the
82 fine fraction of an alpine soil. We differentiate between the physical and chemical processes that
83 drive the production of the scree slope sandy layers and of the alpine soil profile, by comparing their
84 respective grain size distributions, and bulk mineralogy. Furthermore, we examine the potential
85 influence of Sahara and locally sourced aeolian dust accretion on the alpine soil profile by comparing

Commented [A4]: Hematite is an iron oxide, so the phrase has been changed accordingly

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86 the sedimentological, mineralogical, and radiogenic isotope compositions through the application of
87 $^{86}\text{Sr}/^{87}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios between the soil samples and Sahara dust samples collected from the
88 snowpack. We finally measured the magnetic properties of the soil samples and clay mineralogy of
89 bottom and topsoil layers, to assess the potential for weathering of clay minerals and iron oxides
90 within Mount Olympus periglacial zone. Understanding the sources of parent materials and soil
91 formation processes between contrasting geomorphological settings is a fundamental step towards
92 defining the postglacial paleo-environmental history of Mount Olympus alpine landscapes that
93 followed pronounced shifts of the regional climate.

Commented [A7]: We changed the sentence and replaced the 'potential for pedogenesis' with 'potential weathering of iron oxides', which corresponds directly to the analyses undertaken and to the overall goal of the manuscript

95 2 BACKGROUND

96 2.1 Mount Olympus glacial history

97 Mount Olympus is the highest mountain in Greece, rising 2918 m above the northwest coastline of
98 the Aegean Sea (Figure 1a). It is a precipitous massif with a circular shape composed of Triassic to
99 Cretaceous metacarbonates, uplifted along a frontal fault that runs parallel to the present-day
100 shoreline. Mount Olympus is exhumed from the silicate crystalline bedrock, which dominates the
101 lithology of Pieria Mountains (granites, ophiolites) to the north and east, and Mount Olympus
102 granites to the west (Figure 1B). High uplift rates along with successive Quaternary glaciations have
103 created the present-day rugged terrain. The deglaciation of Mount Olympus since the Last Glacial
104 Maximum (LGM), between 28 and 24 ka BP (Allard et al., 2020), triggered the rapid retreat of an ice
105 cap that was covering the summit area and extended down to elevations of ≈ 1800 m (Kuhlemann et
106 al., 2008). The post-LGM glacier retreat was intercepted by a glacier re-advance phase at ≈ 15 ka BP
107 that was limited at the highest cirques above 2200 m at ≈ 12.5 ka BP (Styllas et al., 2018). This latter
108 phase of glacial expanse is traced in both Megala Kazania (MK) and Throne of Zeus (TZ) cirques
109 (Figure 1C). The absence of absolutely dated glacial features between early- and mid-Holocene (9–4
110 ka BP) in both cirques suggests reduced glacial activity, whereas Late Holocene (4 ka BP to present)
111 glacier advances have been observed only in the MK cirque (Figure 1C). These include a terminal

112 moraine stabilization phase at \approx 2.5 ka BP followed by a smaller expansion of the MK glacier at the
113 beginning of the Little Ice Age (LIA) at \approx 0.6 ka BP (Styllas, 2020). Late Holocene glacier advances in
114 the MK cirque lack similarly dated glacial landforms in the TZ cirque, but we cannot rule out the
115 possibility that the Late Holocene climatic shifts towards glacial conditions triggered an
116 intensification of glacial and periglacial processes, which in turn affected the late Holocene landscape
117 evolution, scree slope aggradation and alpine soil production.

118

119 **2.2 Climate**

120 The contemporary maritime conditions and the steep relief of Mount Olympus result in intense
121 precipitation and temperature altitudinal gradients, with the highest peaks constituting an
122 orographic and climatic barrier between the eastern (marine) and western (continental) sides (Figure
123 1b, Styllas and Kaskaoutis, 2018). The climate in the coastal zone is typically Mediterranean, whereas
124 at higher elevations (1000–2200 m), the climate attains sub-Mediterranean characteristics with
125 average annual precipitation of 1300 mm (Styllas et al., 2016). In the alpine zone above the tree line
126 (2400 m), the climate is characterized by temperate conditions with annual precipitation above 2000
127 mm and average annual temperatures between 0 and 1.5 °C (Styllas et al., 2016). The periglacial
128 activity in the Mount Olympus alpine zone is likely still active today, as it is situated just above the
129 lower limit of the regional permafrost zone (2700m) of the southern Balkan peninsular (Dobiński,
130 2005).

131

132 **2.3 The Plateau of Muses**

133 The Plateau of Muses (PM) is a planar depositional surface located at an elevation of 2600 m with a
134 surface area of 1 km². It resembles a typical alpine meadow, partly covered by alpine grass
135 vegetation that shares similar characteristics with plateaus found in the high Balkan Mountains and

136 the European Alps. The PM is bounded to the south by the TZ cirque lateral moraine ridge and by
137 several gentle-sloping glacially eroded peaks along its northern, eastern, and western margins (Figure
138 1C). The formation of the plateau has resulted from the combined action of glacial scouring and
139 carbonate bedrock dissolution. Its low relief in combination with the circular shape suggest a doline
140 type karstic depression that is filled with glacial till, overlain by colluvial sediments (slope wash)
141 transported from the adjacent slopes. The surface layer of the PM sedimentary sequence comprises
142 a developed soil sequence with variable thickness (30–50 cm) that overlies a layer of outwash sand
143 and fine gravels and/or fragmented till boulders, and exhibits brown-red to yellow color hues, which
144 in the Munsell color scale range between 7.5 and 10 YR (Table 1). Alternating patches of alpine grass
145 vegetation and hummocky soil pans in the center of the plateau are indicative of periglacial activity
146 and cryoturbation. Other periglacial features such as solifluction-terraced stripes below the bare
147 bedrock of the surrounding summits are tentatively considered to have formed during the Late
148 Holocene cold stages, during the observed expansion of small glaciers in the MK cirque.

149

150 **3 MATERIALS AND METHODS**

151 **3.1 Erosional products and alpine soil sampling**

152 To adequately address the question of the relative balance between aeolian dust accretion and local
153 erosion of moraines and scree slopes to the development of the alpine soil on Mount Olympus
154 periglacial zone, a wide range of methods were employed and involved the analyses of 21 discrete
155 soil and sediment samples retrieved along a transect of decreasing hillslope energy and erosional
156 power (Figure 2). Five samples (n=5) were retrieved from clast-free sandy horizons interbedded in
157 the relatively young (Late Holocene) MK and older (early Holocene) TZ stratified scree slopes, and
158 sixteen (n=16) were sampled from the PM soil sequence at 2-cm intervals (Table 1). The specific
159 experimental setting was selected to evaluate the impact of physical weathering on providing the
160 base material for the development of the PM soil. We only sampled naturally exposed clast free
161 sandy layers found within the scree slopes of MK and TZ. We considered that these layers share

Commented [A8]: We elaborate on the study's experimental design and explain the selection of the specific sampling sites.

162 similar textural, mineralogical, and geochemical characteristics with the PM soil basal horizon, which
163 lies on a layer of outwash sand and gravels. Luckily, we were able to retrieve the samples from two
164 distinct interbedded clast free sediment layers within the TZ scree slope after a torrential rainfall
165 event that opened a deep erosional trench in the scree below the rockwall and reached the basal till
166 layer (Figure 2). The scree slope in the MK is regularly eroded and scoured from a perennial snowfield
167 that is retreating by the end of the summer season, and this made the sampling of distinct soil-
168 sediment horizons straightforward. We manually excavated only one pit for high resolution soil
169 sampling and considered that due to the very small surface area of the surficial soil apron within the
170 PM catchment (0.06 km²), the specific profile is representative of the PM soil development. We
171 selected a location in the center of a circular soil-sediment pan that was free of vegetation, surface
172 carbonate fragments (Figure 2). After sampling, the pit was closed and refilled with the excavated
173 material in accordance with Mount Olympus National Park directions. In locations with long lasting
174 snowpack, we observed a humic A horizon, but since these locations host several endemic flower
175 species, the Management Unit of Mount Olympus National Park, did not grant permission to
176 excavate a soil pit in these sensitive sites. The PM soil samples were additionally subjected to
177 microscopic and radiogenic isotope analyses and magnetic measurements to investigate the
178 potential chemical alterations processes during PM soil development. Mineralogical and radiogenic
179 isotope analyses were also performed in two (n = 2) samples of aeolian dust that were deposited on
180 the PM snowpack during the spring seasons of 2018 and 2022. The long-range aeolian dust transport
181 episodes occurred on March 22–24, 2018, and March 16–18, 2022. The synoptic conditions of these
182 distinct episodes show that the dust emissions traveled to Mount Olympus from the Sahara Desert
183 and left an orange hue on the snowpack, which later in the spring season formed distinct layers in
184 the snowpack (Figure 3). We therefore consider the samples collected from the PM snowpack as
185 representative of Sahara dust accretion in Mount Olympus alpine soils.

186

187 **3.2 Grain size analyses**

188 The soil samples were transported to the lab, wet sieved through a 3.5-mm sieve, and treated with
189 30% hydrogen peroxide (H₂O₂) at 70 °C for 12 h to remove organic matter. The H₂O₂ treatment was
190 repeated three times until the samples were completely bleached and all organic matter was
191 degraded. The samples were washed with distilled water and analyzed with a Mastersizer 3000 laser
192 diffraction particle-size analyzer to define the bulk grain size distributions of the sand, silt, and clay
193 fractions. The samples were run through the automated dispersion unit and sodium
194 hexametaphosphate solution (Calgon) was added as dispersion factor. Statistical analyses of the grain
195 size distributions and derivation of the clay, silt, and sand fractions were realized with MATLAB Curve
196 Fitting Lab (CFLab), which performs curve fitting on sediment grain size distributions using the
197 Weibull probability distribution function (Wu et al., 2020).

198

199 3.3 Mineralogy

200 Identification of the mineral phases of the soil and aeolian dust bulk samples was achieved through
201 X-ray diffraction (XRD, Philips diffractometer PW1800, Co radiation at 40 kV and 40 mA), and two
202 samples from the top and base of the PM soil profile (PM1 and PM15) were additionally analyzed for
203 their clay (<2 µm) mineralogy through ethylene glycolation and heating for 2 h at 550 °C. The PM soil
204 samples semi-quantitative composition of the main mineral phases (e.g., quartz, feldspar,
205 plagioclase, micas, calcite) was determined using MAUD-Material Analysis software applied for full
206 pattern Rietveld refinement (Lutterotti et al. 2007) and is expressed as weight percent (wt %)
207 concentrations.

208

209 3.4 Petrographic, magnetic, and isotopic analyses

210 Additional analytical methods were applied only to the PM soil samples to assess the potential sources
211 of soil-forming material, pedogenesis, and chemical weathering. The fabric configuration of the PM
212 alpine soil was explored through scanning electron microscopy–energy dispersive spectrometry
213 (SEM–EDS) analyses (JEOL JSM-840A equipped with an INCA 250; Oxford) with a 20-kV accelerating

Commented [A9]: This has been a typographical error in the previous versions of the manuscript. We sieved the soil and sediment samples through a 3.5mm sieve. We also include an Excel file with the results of the Mastersizer 3000 grain-size analyses to provide further evidence for this error.

Commented [A10]: Typo has been corrected

214 voltage and 0.4-mA probe current. Backscattered electron images (BSE) enabled us to detect the
 215 shapes of different minerals, and the physical weathering features of specific grains, whereas with
 216 the EDS analysis we examined areas of different chemical composition within the same soil
 217 aggregates.

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218 We additionally explored the existence of ferromagnetic components and the potential for
 219 secondary iron oxides formation in the PM soil profile through magnetic susceptibility
 220 measurements. The discrete samples were packed in cubical plastic boxes ($2 \times 2 \times 2$ cm) and weighed
 221 before the measurements. Volume-specific magnetic susceptibility measurements were performed
 222 using both a Bartington dual MS2B sensor at low and high frequencies of 0.465 and 4.65 kHz. The
 223 results are expressed as mass-specific magnetic susceptibility (χ ; 10^{-8} m³/kg). During the measuring
 224 procedure, every sample was measured at least three times and the average value was assigned as
 225 the final measurement. For each sample, two air measurements were performed before and after
 226 sample measurement. The frequency-dependent susceptibility (χ_{FD} ; %) was calculated according to
 227 Dearing et al. (1996):

Commented [A12]: "Potential for pedogenesis" is replaced and further explained.

$$228 \quad \chi_{FD}\% = \frac{100(\chi_{LF} - \chi_{HF})}{\chi_{LF}} \quad (1)$$

229 where χ_{LF} , χ_{HF} , are the magnetic susceptibility at low and high frequency, respectively. Samples PM16
 230 and PM15, which were considered as more representative of the PM soil regolith boundary, were
 231 additionally subjected to thermomagnetic analysis to define the origin of the ferromagnetic particles
 232 at the base of the PM soil. Measurements of continuous thermomagnetic curves (K-T curves) at low
 233 and high temperature were realized with the furnace CS3 of the AGICO MFK1-FA susceptibilimeter.

234 The potential sources of the PM soil and aeolian dust were evaluated through their Sr and Nd
 235 isotopic ratios. Isotopic measurements were performed at the University of Arizona TIMS laboratory
 236 following the procedure in Conroy et al. (2013) on soil samples. Samples were not spiked and
 237 dissolved in mixtures of ultrapure Hf-HNO₃ acid. Elemental separation of dissolved samples was
 238 carried out in chromatographic columns via HCl elution in a clean laboratory environment.
 239 Conventional cation columns filled with AG50W-X4 resin were used for Sr and REEs separation and

240 anion columns with LN Spec resin for Nd separation following Ducea et al. (2020). Sr cuts were
 241 loaded onto Ta single filaments and Nd cuts onto Re filaments. $^{87}\text{Sr}/^{86}\text{Sr}$ and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios (Table
 242 3) were measured on a VG Sector 54 thermal ionization mass spectrometer (TIMS) fitted with
 243 adjustable 1011 Ω Faraday collectors and Daly photomultipliers. NBS SRM 987 Sr standard and La
 244 Jolla Nd standard were analyzed during the samples run to ensure the performance of the
 245 instrument and to perform some minor correction on the final reported ratios.

246

247 3.5 Erosional potential and aeolian dust accretion proxies

248 The erosional potential of the three sampling sites, which are distanced along a 2km transect was
 249 derived from field estimates of the vertical height of the MK and TZ rocky headwalls and their scree
 250 slopes. We evaluated the plateau and scree slope energy distribution and maturity stage from the
 251 dimensionless ratio between the vertical height of the scree slope (Ht) to the vertical height of the
 252 headwall (Hc) following Statham (1976) (Figure 2).

253 To assess the potential contribution of distal and local aeolian dust inputs in the PM soil we used
 254 the contents of quartz (wt. %). The source of quartz can be local, from the Pieria Mountains silicate
 255 bedrock and from the granites to the west of Mount Olympus, or can be transported during Sahara
 256 dust episodes, as evidenced from the XRD analyses of the PM snowpack samples, which are in line
 257 with Sahara dust samples from the Pyrenees, the European Alps, and the Carpathian Mountains that
 258 contain high amounts of quartz (e.g., Rellini et al., 2009; Rodriguez-Navarro, 2018; Marmureanu et
 259 al., 2019). Herein, we cannot exclude the possibility of quartz release from the local bedrock through
 260 periglacial erosion, but the amount of quartz released from local bedrock dissolution is expected to
 261 be small, wt.% concentration of the insoluble residue from carbonates in Greece is less than 1%
 262 (MacLeod, 1980; Kantiranis, 2001; Kirsten and Heinrich, 2022). Therefore, it is reasonable to consider
 263 quartz (wt. %) as a reliable proxy of aeolian dust accretion.

264 We selected the ϵ_{Nd} ratio as a second independent proxy particularly of Sahara dust accretion
 265 in the PM soil. We did not use the Sr ratio ($^{87}\text{Sr}/^{86}\text{Sr}$) as it can be impacted by the dissolution of

Commented [A13]: The paragraph was moved to methods according to Reviewer's 1 suggestion.

Commented [A14]: Responding to Reviewer 1 comment that quartz is not a reliable proxy of aeolian dust, we provide references that the potential contribution of quartz from the local bedrock is minimal and that quartz has been found to dominate Sahara dust deposits on the snowpack of several mountain ranges.

266 carbonate particles and replacement of Ca by Sr during pedogenetic alteration of the PM soil (e.g.,
 267 Shalev et al., 2013). Sr isotopic distributions of PM soil can be further complicated by the accretion of
 268 sea-salt Sr through orographic precipitation (Kurtz et al., 2001). Rain is not a significant source of Nd,
 269 so the addition of rainwater and snow should not affect the Nd isotopic composition of the aeolian
 270 dust, so ϵ_{Nd} is buffered against these changes (Kurtz et al., 2001). We estimated the fraction of
 271 Sahara dust from the ϵ_{Nd} ratios of the PM soil, the dust deposited on the snowpack, and the local
 272 bedrock following the method by Kurtz et al (2001):

$$f = \frac{(\epsilon_{Nd} \text{ PM soil} - \epsilon_{Nd} \text{ bedrock})}{(\epsilon_{Nd} \text{ Sahara dust} - \epsilon_{Nd} \text{ bedrock})} \quad (2)$$

275 As we had not obtained direct Sr and ϵ_{Nd} values from Mount Olympus bedrock, we used the value of
 276 the basal sample PM 16, which is dominated by bedrock derived calcite and falls in the same value
 277 range with basin-average values of terrestrial, coastal, and marine sediments deposited in the
 278 Aegean Sea (Weldeab et al., 2001).

Commented [A15]: Sample PM 16 is sandy and gravely and is characterized by similar values with the Aegean Sea sediments (see Figure 10), so we considered the average Sr- ϵ_{Nd} values representative of the regional carbonates.

280 4 RESULTS

281 4.1 Alpine soil formation across a hillslope energy gradient

282 According to Statham (1976), Ht/Hc values above 0.4 characterize a mature scree slope, which is the
 283 case for the TZ (0.6) but not for the MK (0.3) scree slope, and this reflects the older deglaciation age
 284 of the TZ cirque (≈ 12.5 ka BP, Section 2.1). The MK scree slope is deposited behind the LIA moraine
 285 (Figure 1C), so that the most recent deglaciation processes (≈ 0.6 ka BP to present) has resulted in
 286 immature scree development. Conversely, the low-relief, low-erosion PM acts as a long-term
 287 depocenter of slope wash and detrital (aeolian and bedrock derived through freeze-thaw action)
 288 sediments. For this low-energy setting, we believe that minor colluvial contributions, cryoturbation,
 289 aeolian dust accretion, fine earth translocation, and post-depositional mineral alteration are the
 290 major drivers of PM soil production. The irregular boundary between the base of the PM soil and the

Commented [A16]: Speculative expressions have been changed to affirmative ones.

Commented [A17]: Sentence has been deleted.

291 underlying regolith composed of glacial till and outwash gravels, is indicative of cryoturbation, while
 292 observations of late-season soil freezing and waterlogging (Figure 4) provide permissive evidence
 293 that PM soil development is disturbed by cryogenic processes. The energy gradient along the
 294 contrasting environments impacts the soil color. The PM soil basal layer overlying the regolith shares
 295 similar color characteristics with the MK samples and with the TZ upper sediment horizon, which
 296 have grey to olive green hues (Munsell dry color 2.5–5 Y; Table 1). Conversely, the lower clast-free
 297 sediment horizon of the TZ scree shares similar Munsell dry color characteristics with the PM soil,
 298 characterized by red-brown to yellow hues (7.5–10 YR, Table 1), suggesting that these soil samples
 299 are more oxidized and are undergoing pedogenetic alterations.

300

301 4.2 Grain size variation

302 The interactions between slope processes, colluvial sediment transport, and aeolian sub additions
 303 result in polymodal grain size distributions that display different shapes among MK, TZ, and PM soils.
 304 Five grain size modes (M1 to M5) were mathematically derived from the application of the CFLab
 305 curve-fitting algorithm. Fitting degrees were >99% and fitting residuals were <0.1%, indicating
 306 excellent fits for the raw grain size distribution curves (Figure 5 A, B, C, and D). The fine-earth (clay
 307 and silt) fractions resemble grain size modes M1 and M2 with respective mean grain sizes of ≈ 2 and
 308 ≈ 4 μm and M3 with mean grain sizes between 10 and 30 μm . The sand fraction is composed of two
 309 modal sub-populations: a fine-sand-grain size mode (M4, mean grain size ≈ 80 μm) and a coarse
 310 sand-grain size mode (M5, mean grain size 440 μm) (Figure 5 E, F, G, and H; Table 1). The production
 311 of coarse sand is transported to the respective interbedded sediment horizons by rockfall activity and
 312 colluvial processes, or in the case of the low-sloping PM, through slope wash. The fine sand (M4)
 313 subpopulation was not traced in the PM soil samples, and this can be linked to either selective
 314 entrainment of M4 or to distortion of the MK and TZ grain size curves and truncation of the coarser
 315 modes (Garzanti et al., 2009).

Commented [A18]: As previously stated, there has been a typo in the previous versions of the manuscript, as the range of the sieved sediments and soil samples was 3.5mm.

Commented [A19]: Following the comments of Reviewer 1, we removed the sentence "The presence of fine (M4) and coarse (M5) sand is linked to frost weathering of the carbonate bedrock." as we do not have adequate evidence to support this.

316 In addition to the distinct color variations, the contrasting slope-energy distribution between the
317 MK, TZ scree slopes and the PM depositional environments also defines their textural compositions.
318 Sediment horizons developed on the surface of the MK scree slope contain higher amounts of sand
319 (\square 90%) and lower amounts of silt and clay (\square 10%) compared with their TZ counterparts (\square 75% and
320 \square 25%), implying that the dominance of sand in the sediment horizons of the scree slopes derives
321 from freeze-thaw and colluvial activity. The coarse-sand content (M5) of the PM soil basal layer is 6%
322 but is lower within the solum (2%–3%), suggesting either reduced periglacial activity and/or low
323 transport capacity of erosional products from the catchment through slope wash processes during
324 the PM soil formation (Table 1; Figure 2, lower graphs).

325 The grain size distribution curves of the PM soil present a significant change in shape between
326 soil depths of 14 and 16 cm, which is characterized by a 15% reduction of the clay and very-fine-silt
327 fractions (M1 and M2) and by a similar increase of silt contents (Figure 5A and B). This sharp textural
328 differentiation was not supported from field observations, where the solum appeared homogenous
329 without distinct pedogenetic horizons and without any visual evidence of an erosional layer (Figure
330 6A), but it is supported by changes in the soil color. The samples above a soil depth of 14–16 cm
331 exhibit red to brown hues (7.5 YR), whereas the samples below this layer have more yellow-red (10
332 YR) hues (Table 1). We also observed clay coatings in sparse secondary carbonates (calcretes) along
333 the lower part of the PM soil profile, which we interpret as evidence of soil mixing and downward
334 translocation of dissolved Ca and secondary calcite precipitation at the base of the soil profile. Based
335 on these observations, we partitioned the PM soil profile in two horizons: an upper Bw horizon
336 between 0 and 14 cm with red to brown hues, low clay (\square 25%), and high silt (\square 75%) contents, and
337 a lower illuvial Bt horizon between 14 and 32 cm with higher (\square 40%) clay contents, a yellow-red hue,
338 and smaller amounts (\square 5%) of sand compared with the overlying Bw horizon (Figure 6).

339

340

Commented [A20]: Based on these observation we conclude that the sand populations are highest in the immature scree slope interbedded sediment layers and thus sand can be considered representative of periglacial erosion.

341 4.3 Soil and aeolian dust mineralogy

342 XRD analysis of the bulk samples reveals a mineralogy that substantially differs between the MK,
343 TZ, and PM soils and, like the soil colorization and textural variations, follows the erosional slope
344 gradient. The most dominant mineral phase in the clast-free material of the MK and TZ soils is calcite.
345 Other minerals identified include dolomite along with quartz and micas. Conversely, the bulk
346 mineralogical composition of PM soil exhibits a richer matrix of minerals that includes quartz, chlorite
347 and mixed layer clays, mica, potassium feldspars, and plagioclase (Figure 7). Calcite is dominant
348 (\square 50%) only in basal sample PM16 (Figure 6; Table 2). Quartz, clays, and mica are the most
349 dominant mineral phases in the PM soil (\square 80%) with low values in basal sample PM16, whereas
350 plagioclase, K-feldspar, and mica represent the remaining 20% (Table 2). Semi-quantitative analysis
351 of the clay mineralogy of two samples retrieved from the surface of the Bw horizon and the base of
352 the Bt horizon (samples PM1 and PM15) revealed high concentrations of smectite and kaolinite
353 (80%) and low contents of chlorite and illite. Surface sample PM1 contains 45% smectite and 35%
354 kaolinite, whereas basal sample PM15 has higher smectite (65%) and lower kaolinite (25%) contents
355 (Table 2).

356 From the comparison of the XRD spectra (Figure 7), it is obvious that the bulk mineralogy of the
357 PM soil matches that of the Sahara dust samples. Both Sahara dust samples show the presence of
358 clay minerals, quartz, mica, calcite, plagioclase, K-feldspar, and dolomite. The detected mineral
359 phases are typical of Saharan dust deposited in Europe during both dry- and wet-deposition (red
360 rains) events (Scheuven et al., 2013). Additionally, recent studies of Saharan dust wet deposition in
361 the Iberian Peninsula also indicated the presence of Fe-Ti oxides, such as goethite and hematite, and
362 of Ti oxides, such as rutile (Rodriguez-Navarro et al., 2018), but these were not depicted from our
363 XRD analyses. Despite their overall XRD spectral similarity, a pronounced difference between the
364 contemporary Sahara dust and PM soil samples is the presence of calcite and dolomite in the dust
365 samples and their near absence from the PM soil profile (Figure 7). The smooth and low intensity

366 peaks for calcite and dolomite at 29.43 and 30.7 2 θ in surface sample PM1 indicate the partial
367 removal of calcite, whereas similarly subdued peaks in basal sample PM16 denote near complete
368 decalcification of the solum (Figure 7).

369

370 **4.4 Magnetic susceptibility of PM soil**

371 The magnetic susceptibilities of the PM soil bulk samples were measured to provide insight into the
372 ferromagnetic components of the PM soil and their potential alterations. Overall, the low-frequency
373 magnetic susceptibility (χ_{lf}) is higher in the lower Bt horizon, with average values for samples PM8–
374 PM16 of $55 \times 10^{-8} \text{ m}^3 \text{ kg}^{-1}$, and lower χ_{lf} values in the Bw horizon, with average values for samples
375 PM1–PM7 of $36 \times 10^{-8} \text{ m}^3 \text{ kg}^{-1}$ (Figure 8A). Similar value ranges were measured for the high-
376 frequency magnetic susceptibility (χ_{hf}). The estimated values of frequency-dependent (χ_{FD})
377 susceptibility presenting a wide range of values ranging between 0% (sample PM13) and 14%, with
378 significantly higher values in the Bw horizon (Figure 8B). According to Dearing (1999), high χ_{FD} values
379 (>10%) are indicative of the presence of superparamagnetic Fe oxide nanoparticles (< 0.05 μm),
380 suggesting a higher amount of fine ferrimagnetic grains in the surface horizon Bw, which potentially
381 can be of detrital (aeolian and/ eroded bedrock) origin.

382 The mineral phases responsible for the magnetic enhancement of the Bt horizon were deduced
383 from high-temperature magnetic susceptibility measurements performed during a single heating–
384 cooling cycle to 700 °C (Figure 8C). We estimated the Curie temperature (T_c) of samples PM16 and
385 PM15 to examine the potential existence of superparamagnetic ultrafine particles in the base of the
386 PM soil profile, which is in contact with the regolith. The thermomagnetic analysis of sample PM16
387 failed completely, likely due to its high calcite content and absence of magnetic phases. On the other
388 hand, sample PM15 resembling the soil-regolith lower boundary, revealed a uniform χ -T behavior
389 that is indicative of the dominance of two magnetic phases (Figure 8C) – one with T_c , or
390 transformation temperature, between 260–320 °C, probably maghemite, and a second one around

391 600 °C, which is typical for oxidized magnetite (Jordanova et al., 2022). Since the nano-sized
392 pedogenic magnetite is unstable upon heating (Dunlop and Özdemir, 1997), the identified oxidized
393 magnetite suggests that weak pedogenetic production of ferromagnetic components occurs in the
394 base of PM soil profile.

395

396

397 **4.5 Radiogenic isotopes**

398 More information on the provenance of the PM-soil-forming material was derived from the
399 radiogenic isotope analysis of the soil samples and of the 2018 Sahara dust sample. The $^{87}\text{Sr}/^{86}\text{Sr}$
400 values of PM soil samples range from 0.71437 to 0.72071 and the ϵ_{Nd} values from -7.75 to -9.80
401 (Table 3). Overall, the PM soil $^{87}\text{Sr}/^{86}\text{Sr} - \epsilon_{\text{Nd}}$ cluster together apart from sample PM16, which has the
402 lowest value of the PM soil $^{87}\text{Sr}/^{86}\text{Sr}$ ratio (Figure 9). The analyzed Sahara dust sample exhibits
403 $^{87}\text{Sr}/^{86}\text{Sr}$ value of 0.71272 that falls within the lower range of North African dust sources between
404 \square 0.71200 and 0.74000 (Erel and Torrent, 2010; Grousset and Biscaye, 2005). The Sr isotopic ratio of
405 the Sahara dust sample shows potential mixing with rainwater and local sea salt aerosols during the
406 March 2018 wet deposition event but also with other European aerosol sources, which is validated
407 by the fact that the dust plume of the March 2018 travelled over Europe before it reached Mount
408 Olympus (Figure 3A). The Sahara dust sample has an ϵ_{Nd} value of -6.80. Plotting the $^{87}\text{Sr}/^{86}\text{Sr}$ and ϵ_{Nd}
409 measurements against literature values from terrigenous, coastal, and marine sediments from the
410 Aegean Sea region (Weldeab et al., 2002) reveals an isotopic similarity between the Sahara dust and
411 of sample PM16 with these sediments (Figure 9). A reasonable interpretation of this observation
412 comes from the fact that basal sample PM16 resembles more the soil regolith and plots close to the
413 contemporary and Holocene values of Aegean Sea terrestrial, coastal and marine sediments. The
414 $^{87}\text{Sr}/^{86}\text{Sr}$ values representing the PM soil regolith show similar values with the Aegean Sea terrestrial
415 and marine sediments and likely represent a mix of Sahara dust with Mesozoic and Cenozoic bedrock
416 carbonates, which are overall characterized by low $^{87}\text{Sr}/^{86}\text{Sr}$ values of <0.70800 (Capo et al., 1998;

417 Frank et al., 2021). The two subclusters of PM soil samples have more radiogenic values compared
418 with those of the rest of the samples and clearly correspond to Bw and Bt horizons. The increasing
419 silt contents towards the surface of the PM soil profile (Figure 6, Table 1) occur with $^{87}\text{Sr}/^{86}\text{Sr}$ and ϵ_{Nd}
420 values towards more crustal values (color variation in Figure 9) that are representative of the central
421 Sahara province. Therefore, the increases in the silt fraction within the PM soil profile can be directly
422 linked to increases in Sahara dust accretion. This is further supported by range of the silt fraction
423 mean grain size between 14 and 30 μm (Table 1), which is similar to those for modern Sahara dust
424 deposits from Crete, which range 4–8 μm and 16–30 μm (Mattson and Niéhlen, 1996; Goudie and
425 Middleton, 2001). In terms of Sahara dust provenance fingerprinting, the $^{87}\text{Sr}/^{86}\text{Sr}$ and ϵ_{Nd} values of
426 the PM soil samples fall within the range (1σ) of the central North African dust source area, which
427 broadly involves the Bodele depression (PSA2; Jewell et al., 2021).

428

429 5 DISCUSSION

430 5.1 PM soil parent material

431 The mineralogical (XRD) analyses, show that calcite is the dominant mineral phase of MK and TZ
432 interbedded sandy sediment and PM basal layers (Figure 7, lower XRD diagrams TZ01 and Table 2
433 sample PM 16), which in the periglacial environment of Mount Olympus is expected to dissolve
434 slowly (e.g., Gaillardet et al., 2019) and produce an insoluble residue that comprises the PM soil
435 parent material. MacLeod (1980) analyzed the mineral composition of the insoluble residue of
436 carbonates from western Greece and defined a mineralogical suite of quartz, kaolinite, and mica
437 (illite). Kantiranis (2001) studied the carbonate rocks of northwestern Greece and found insoluble
438 residue $\approx 1\text{wt.}\%$ consisting mainly of micas, quartz, hematite, chlorite, feldspars, and amphibole,
439 whereas the insoluble residue of carbonate basement rocks from Crete also resembles $\approx 1\text{ wt.}\%$ of
440 the whole rock samples and is composed of a sandy loam matrix rich in quartz, plagioclase (albite),
441 and mica (illite) (Kirsten and Heinrich, 2022). Thus, the dissolution of the local carbonate parent

442 material within the interbedded sediment layers and in the basal layer of PM soil, can release very
443 small quantities of bedrock-derived impurities such as quartz, plagioclase, illite, and kaolinite that are
444 incorporated in the solum, but cannot explain the ≈ 30 cm thick PM soil mantle and ≈ 60 cm thickness
445 of the layers interbedded in the scree slopes.

446 It has also been proposed that clay in terra rossa soils can derive from isovolumetric
447 replacement of calcite to authigenic clays across a metasomatic front, but this mechanism requires
448 significant input of aeolian dust to provide essential elements such as Al, Si, Fe and K for clay
449 formation (Merino and Banerjee, 2008). Even though we did not estimate the dissolution rate of
450 Mount Olympus bedrock metacarbonates and the elemental composition of the insoluble residue,
451 we consider that the fine earth (silt and clay) contents of MK and TZ interbedded layers, which
452 average 10% and 25%, respectively, cannot be derived only by carbonate dissolution and/or by
453 isovolumetric replacement of calcite. Küfmann (2008), Krklec et al. (2022) and Ott et al. (2023)
454 propose carbonate bedrock dissolution rates between ≈ 0.23 , 0.15 and 0.4 cm/ka respectively, which
455 for the postglacial (12.5 ka BP to present) alpine soil formation on Mount Olympus imply ≈ 5 cm of
456 carbonate loss to soil formation, a value too low to explain the observed thickness of MK, TZ,
457 interbedded layers and PM soil as a result of residual clay accumulation alone. Our direct
458 observations of episodic Sahara dust deposition on the snowpack of Mount Olympus (Figure 3)
459 provide undisputable evidence of Sahara dust accretion on PM soil. The relative contribution of local
460 dust from moraines, outwash plains and from silicate bedrock formations in the vicinity of Mount
461 Olympus is estimated in the following section, but irrespective of the relative dust sources (Saharan
462 and local), the high-energy erosive regime of Mount Olympus alpine critical zone intercepts the
463 formation of extensive aeolian dust mantles, like the one found on the stable Plateau of Muses. We
464 thus suggest that the production of silt, and clay in the PM soil basal layer, partly reflects the
465 contribution of mechanically produced sandy and fine earth carbonate debris and its dissolution
466 products, which together with aeolian dust accretion, comprise the parent materials for the PM soil
467 production.

Commented [A21]: Herein we discarded the section L 480-488 of the previous version of the manuscript as we did not have enough evidence to back up our claims of elemental enrichment and secondary clay formation, in agreement with the comments of Reviewer 1.

468

469 **5.2 Relative contributions of aeolian dust inputs**

470 Studies on terra rossa soils in Greece, with typical bimodal grain size distributions consisting of clay
471 and silt subpopulations with grain size ranges of 2–4 and 10–40 μm , respectively, ascribe the clay
472 fraction, which is rich in illite and kaolinite, to the limestone residue, and the silt fraction, which is
473 made up entirely of quartz, to long-range aeolian transport from variable sources (Russel and Van
474 An del, 2003). In line with this notion, we considered the quartz wt. % content in the solum, as a
475 proxy for aeolian dust in general and not exclusively of Sahara dust. The rounded shape of quartz
476 grains observed in SEM images (Figure 11D), provides supplementary evidence for the aeolian
477 transport of quartz grains. Furthermore, we consider that the neodymium-derived mass fraction (f),
478 solely a proxy of Sahara dust accretion in the PM soil. This is supported by the high statistical
479 correlation between the silt fraction (M3) with the ϵ_{Nd} -derived f fraction ($R^2= 0.73$, $P< 0.001$) and by
480 the similarity of the grain size ranges between the silt fraction and the modern Sahara dust deposits.
481 The mass fraction (f) of Sahara-dust-derived ϵ_{Nd} was calculated based on the highest ϵ_{Nd} value of
482 sample PM 16 and Aegean Sea sediments ($\epsilon_{\text{Nd}} = -5.94$) and on the lowest value of Sahara dust PSA2
483 ($\epsilon_{\text{Nd}} = -13.81$) end members. The ϵ_{Nd} value of Aegean Sea sediments is considered conservative in
484 relation to that of Mount Olympus bedrock due to the mixing of the carbonate bedrock sediments
485 with other sources of silicate bedrock during fluvial transport.

486 The ϵ_{Nd} -based Sahara dust contributions to the PM soil varies between \square 35% and \square 50%
487 (except that of basal sample PM16) (Figure 10). Conversely, the quartz-derived aeolian dust
488 contribution ranges between \square 45% and 65%, shows a relatively small variation with depth and an
489 abrupt increase (\square 25%) from sample PM16 to PM15 (Figure 10). The basal sample PM16 exhibits the
490 lowest contributions of quartz concentration, f ratio values (Figure 10) and silt concentrations (Table
491 1, Figure 6) and is considered an outlier representing the regolith-PM soil mix, which agrees with its
492 distinct color and lowest magnetic susceptibility values. The preservation of quartz in the PM soil

493 profile and especially in the lower Bt horizon requires a mechanism of reduced Sahara dust input
494 and/or loss to weathering, with simultaneous inputs of other quartz-rich-derived dust. A pattern that
495 can explain the lower ϵ_{Nd} -based Sahara dust contributions in the Bt horizon and the near steady
496 quartz contents is a shift in atmospheric circulation patterns that resulted in less-frequent dust
497 transport episodes from north Africa along with steady aeolian quartz accretion from local quartz
498 sources. Aeolian quartz from the silicate bedrock formations of the Pieria mountains, Mount
499 Olympus granites and even from the Katerini alluvial plane (Figure 1) can be deposited on Mount
500 Olympus periglacial zone during periods of regional aridity, associated with thinning of vegetation,
501 desiccation of the Katerini alluvial plane, and immobilization of fine dust grains through convection.

502 Based on the above, we tentatively attribute the $\approx 15\%$ difference between the ϵ_{Nd} -based
503 estimates and the quartz-based estimates to accretion of quartz-rich dust from local sources during
504 the formation of the Bt horizon, considering that the contribution of bedrock derived quartz from
505 the insoluble residue is $\approx 1\%$. From the ϵ_{Nd} -based contributions, we estimate that the Sahara dust
506 accretion to PM soil is between $\approx 35\%$ to 50% , whereas local sources can potentially accrete another
507 $\approx 15\%$. Our estimated aeolian dust accretion $\approx 65\%$ is similar to the one in the North Calcareous
508 Alps, where the local contribution of dust from the periglacial zone of the North Calcareous and
509 Austrian silicate Alps is significant (Küfmann 2008), but our estimated Sahara dust contribution in the
510 PM soil is higher than its respective average contribution (20 – 30%, Varga et al., 2016) in interglacial
511 soils of the Carpathian Basin. We attribute this difference to the closer proximity of Mount Olympus
512 to Sahara Desert than the Carpathian basin. Given that these values are conservative estimates, the
513 aeolian contribution may potentially be higher as, in our calculations, we have not included aeolian
514 transported micas, feldspars, and clays that are integral parts of Sahara dust samples deposited on
515 the snowpack. We thus suggest that the aeolian dust accretion comprises a minimum of $\approx 65\%$ of
516 the PM soil parent material and that carbonate bedrock erosion, and pedogenetic production of
517 detrital clays can potentially contribute another $\approx 35\%$ to the development of PM soil.

518

519 5.3 Pedogenetic alterations

520 An alternative mechanism that can explain the nearly homogeneous depth distribution of quartz
521 (Figure 10) is soil mixing by cryoturbation and subsequent translocation of fine earth particles from
522 the upper Bw to the lower Bt horizon. The mechanism of illuviation does not necessarily cancel the
523 climatic forcing of Sahara dust reduction and increase of local dust inputs during the development of
524 Bt horizon, but rather can act synergistically. For example, a cold and arid climatic phase that
525 immobilizes quartz-rich dust from the Mount Olympus and Pieria mountains piedmonts can also
526 reactivate the periglacial processes on the Mount Olympus alpine critical zone, which in turn
527 enhance scree slope aggradation, colluvial activity, intensification of freeze–thaw cycles, and
528 cryoturbation of the soils. Cryogenically induced translocation of detrital (aeolian and bedrock
529 derived) silt and clays deposited on the surface of the Bw horizon, distorts the textural composition
530 and soil properties and results in massive structures like the one we observed in the PM soil profile
531 (Figure 4).

532 Despite the absence of color difference and of distinct layers in the PM soil, the higher magnetic
533 susceptibility values of the Bt compared to the Bw horizon, can result from the enrichment of
534 ferromagnetic minerals during *in situ* weathering of translocated detrital fine earth particles through
535 pedogenesis (Maher, 2011). However, the overall low values of the frequency dependent magnetic
536 susceptibility ($\chi_{FD} < 10$), point to weak pedogenetic alteration of soil (Dearing et al., 1996), which in the
537 base of PM soil profile occurs through the oxidation of ultrafine (titano)magnetite to maghemite
538 (Section 4.4). The SEM-EDS analyses show the presence of ultrafine Fe-Ti grains throughout the
539 solum (Figure 11B) apart from basal sample PM16 (Figure 11A), which is representative of the PM
540 soil regolith. This is further supported from the EDS chemical composition of the calcite grains in
541 basal sample PM 16 that have TiO₂ weight % concentration <1%. On the other hand, magnetite is
542 found attached to clay minerals of the Bodéle depression surface sediments (Moskowitz et al., 2016),
543 which is the major source of Sahara dust in PM soil (Figure 10). Also, magnetic susceptibility

Commented [A22]: We changed the text that cryoturbation can translocate fine particles and not exclusively clays according to the comment of Reviewer 1.

544 measurements of Sahara dust modern deposits in SE Bulgaria show a low frequency magnetic
545 susceptibility value of $\chi_{if}=97 \times 10^{-8} \text{ m}^3 \text{ kg}^{-1}$ (Jordanova et al., 2013), which is close to the values of Bw
546 horizon ($\chi_{if}=86 \times 10^{-8} \text{ m}^3 \text{ kg}^{-1}$) and Sahara dust episodes are known to transport Fe-Ti oxides in the
547 Mediterranean region (Rodriguez-Navarro et al., 2018). We can therefore ascribe the observed Fe-Ti
548 oxides and (titano)magnetite an aeolian origin, from either the local igneous silicate outcrops, or the
549 Sahara Desert. Collectively these observations imply that the magnetic enhancement of Bt compared
550 to Bw horizon can result from a combination of fine earth illuviation of aeolian transported ultrafine
551 magnetic particles from the Bw horizon to the Bt horizon through cryoturbation and subsequent
552 weak pedogenetic modifications that result to the oxidation of magnetic minerals like
553 (titano)magnetite and can also explain the reddish to yellow color hues.

554

555 5.4 Mineral weathering

556 In addition to the weak pedogenesis of ferromagnetic minerals in the base of the PM soil profile, we
557 assessed the mineral weathering potential of non-magnetic minerals through the clay mineralogy
558 composition of basal and topsoil samples PM15 and PM1. Both samples show the dominance of
559 smectite with lesser contributions by kaolinite, chlorite, and illite (Table 2). High amounts of smectite
560 in alpine soils result from the alteration of detrital chlorite and micas deposited on glacier surfaces
561 and are found in proglacial fields in the European Alps and Rocky Mountains (Egli et al., 2003; Egli et
562 al., 2011; Munroe et al., 2015), so that the 20% difference in smectite concentration (Table 2)
563 between the basal and topsoil layers of PM soil can be partly related to enhanced mineral chemical
564 weathering in the base of the solum. Similarly, kaolinite observed in the XRD profiles of the MK, TZ
565 and Sahara dust samples (Figure 7), can be released from the dissolution of bedrock carbonates, as
566 are the cases for western Greece (Macleod, 1980) and Crete (Kirsten and Heinrich, 2022), but can
567 also form through the alteration of other detrital minerals, such as plagioclase (albite), a process that
568 is common in glacial and periglacial environments (Anderson, 2000). Finally, high smectite and

Commented [A23]: We erroneously used a different citation and we correct this. In their work Anderson et al., 2000 emphasize the alteration of plagioclase to kaolinite.

569 kaolinite contents can also be transported during Saharan dust transport episodes (e.g., Scheuven et
570 al., 2013), but specifically they are representative of the western Sahara dust provenance (PSA 1, Figure
571 10; Rodriguez-Navarro et al., 2018). However, smectite and kaolinite are also found in modern
572 Sahara dust samples deposited in Athens, Greece (Remoundaki et al., 2011). Therefore, we consider
573 that the high (>80%) concentration of smectite and kaolinite in the PM soil clay (< 2 μm) fraction
574 reflects the balance between direct aeolian deposition and *in-situ* weathering of detrital (aeolian
575 and/or bedrock derived) micas and plagioclase, but the respective contributions of aeolian-
576 transported versus that of bedrock-derived clay minerals subjected to post-depositional mineral
577 alterations cannot be defined from the existing data.

578

579 5.5 Relative timing of PM soil development

580 Direct observations suggest that cryoturbation is a fundamental pedogenetic process in the
581 development of PM soil and continues today along with the ongoing accretion of the surficial aeolian
582 silt horizon Bw. The occurrence of seasonal soil freezing and lack of vegetation in the PM polygon
583 centers provide evidence that cryoturbation is active, destroying soil horizons and obscuring
584 pedogenetic and chemical weathering signals. However, magnetic, and mineralogical data indicate
585 the occurrence of weathered Fe-(Ti) oxides such as (titano)maghemite, and the dominance of
586 smectite and kaolinite in the soil basal and topsoil layers, which enable us to conclude that mineral
587 alteration, and pedogenetic modifications of deposited aeolian dust and local erosional products are
588 ongoing processes within the PM soil profile, occurring in tandem with cryoturbation.

589 In the absence of absolute datings that can constrain temporally the processes driving the
590 production of PM soil, we hypothesized on its age based on the conclusions drawn from the
591 contributions of aeolian dust, and the impacts of cryoturbation. We tentatively ascribe the
592 deposition of the base colluvial layer and/or the *in-situ* fragmentation of the regolith's till boulders to
593 the most recent period of glacial activity on Mount Olympus. Based on the glacial record of the MK

Commented [A24]: The presence of clays in the Sahara Dust samples from the snowpack had not been mentioned in the previous version of the manuscript and have changed following the comment of Reviewer 1.

594 and TZ cirques, the best candidates of periglacial activity that have likely resulted in the deposition of
595 outwash sand and gravels postdate the moraine stabilization phases at ≈ 12.5 , 2.5, and 0.6 ka BP.

596 However, there is a 10-ka time span between the Holocene–Pleistocene boundary and the
597 late-Holocene glacial expansions on Mount Olympus. Accepting that PM soil formation began after
598 the moraine stabilization phase at ≈ 12.5 ka BP that was common to MK and TZ cirques, its
599 production rate would be $\approx 3 \times 10^{-5}$ m yr⁻¹ assuming that soil erosion in the low-lying PM has been
600 minimal. This rate is considerably lower than respective soil production rates of Alpine and
601 Mediterranean soils formed over the last 10 ka (Egli et al., 2018; Figure 8). In contrast, by considering
602 a late-Holocene age and that the PM soil development postdates the ≈ 2.5 ka BP moraine
603 stabilization phase, the soil production rate is $\approx 1 \times 10^{-4}$ m yr⁻¹ an estimate that is in better
604 agreement with the soil production rates presented by Egli et al. (2018) for both Alpine and
605 Mediterranean soils. Furthermore, a late-Holocene development of PM soil broadly agrees with soil
606 development patterns in diverse geomorphological environments in Crete (Kirsten and Heinrich,
607 2022). If this scenario is correct, then we can further hypothesize that development of the Bt horizon
608 could have lasted between ≈ 2.5 and 1.0 ka BP, before a recorded phase of intense Sahara dust
609 accretion in Mediterranean that resulted from the combined action of an orbitally induced decrease
610 in solar insolation and of increased aridity over North Africa (Sabatier et al., 2020). This shift could
611 potentially explain the sharp textural boundary between the Bt and Bw horizons and the increasing
612 Sahara dust accretion on the upper Bw horizon. The hypothesized development of the Bw horizon
613 over the past 1 ka could have been disturbed by cryoturbation during the LIA (≈ 0.6 ka BP) glacial
614 expansion in the MK and that continues until today. Ongoing work on Mount Olympus alpine critical
615 zone involves efforts to accurately date the MK and TZ scree interbedded layers and the PM soil
616 profile through Optically Stimulated Luminescence dating that is aided by the high concentrations of
617 quartz in the fine earth fraction, as well as additional geochemical analysis and estimates of the local
618 carbonate bedrock dissolution rates and its residual geochemical composition, in an overall attempt

619 to provide a new continuous record of postglacial alpine landscape evolution in the Mediterranean
620 periglacial zone.

621

622 **6 CONCLUSIONS**

623 In this study, we investigated the local processes that lead to the development of alpine soils on a
624 stable landform on Mount Olympus, considering its regional setting representative of Mediterranean
625 carbonate mountains that became gradually ice-free during the Pleistocene–Holocene transition but
626 that have also been affected by late-Holocene climatic shifts towards glacial and periglacial
627 conditions (Oliva et al., 2018). We discussed the relative contributions of erosion, aeolian dust
628 accretion, and post-depositional pedogenesis and mineral alteration by comparing colluvial sediment
629 layers interbedded in scree slopes with a soil B horizon developed on a regolith composed by slope
630 outwash deposits and fragmented till boulders along a 2km hillslope energy gradient with a
631 northeasterly orientation, which is the main direction of glacial cirque development on Mount
632 Olympus.

633 Overall, our results suggest that soils developed in stable landforms like the PM show signs of
634 weak pedogenesis and contain higher amounts of aeolian dust than locally eroded and chemically
635 weathered products. Aeolian dust from local and Saharan sources is accreted in alpine soils formed in
636 periglacial hummocky polygons of the PM and comprises \square 30%–65% of the soil mass weight. This
637 interpretation matches those of several other studies on aeolian dust accretion in alpine soils (e.g.,
638 Gild et al., 2018; Kaufmann, 2008; Munroe et al., 2015; Yang et al., 2016; Kirsten and Heinrich, 2022)
639 and suggests that aeolian dust is the primary parent soil material on Mount Olympus. The major
640 source of Sahara dust deposited on Mount Olympus is the Bodélé depression, which agrees with
641 observations of accreted dust in Crete (Pye, 1992).

642 In the low-erosional environment of the PM, mineral alteration and weak pedogenetic
643 modifications occur throughout the solum, but their signal is blurred by soil mixing due to ongoing

644 cryoturbation. A sharp textural boundary not visible in the field separates an upper weathered soil
645 Bw horizon from the lower Bt horizon, which is magnetically enhanced and enriched in smectite and
646 kaolinite. Radiogenic isotope systematics, mineralogy, and magnetic susceptibility value range
647 classify the Bw horizon as an aeolian silt layer that was likely formed during a late-Holocene shift of
648 regional atmospheric circulation that resulted in increased Sahara dust accretion in alpine
649 Mediterranean landscapes.

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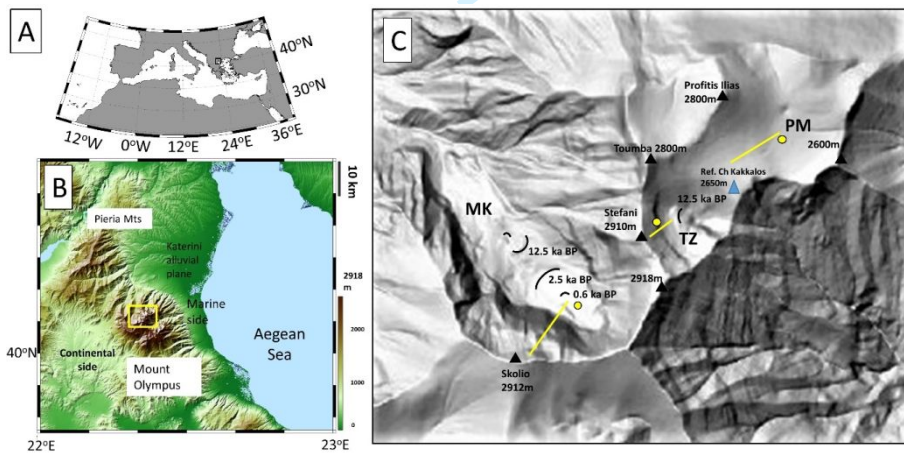
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877 **FIGURES**

878 **Figure 1.** (A) General setting of the study area within the Mediterranean basin. (B) Mount Olympus
 879 alpine domain that is considered in the study (yellow box), with the two respective piedmonts on the
 880 marine and continental sides, the adjacent Pieria Mountains and Katerini alluvial plane (SRTM 90
 881 DEM Model). (C) The highest cirques and plateau of Mount Olympus with the respective locations of
 882 dated moraines (black curved lines from Styllas et al., 2018), the sampling locations considered in this
 883 study (yellow circles) and the geomorphological transects described in Figure 2 (yellow lines). MK:
 884 Megala Kazania cirque, TZ: Throne of Zeus cirque, PM: Plateau of Muses.

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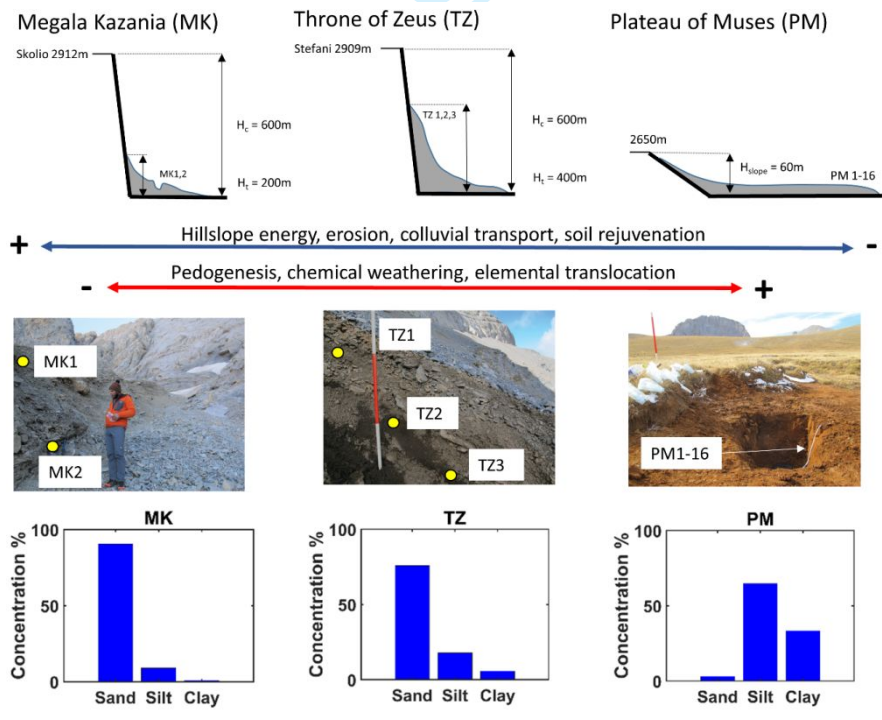


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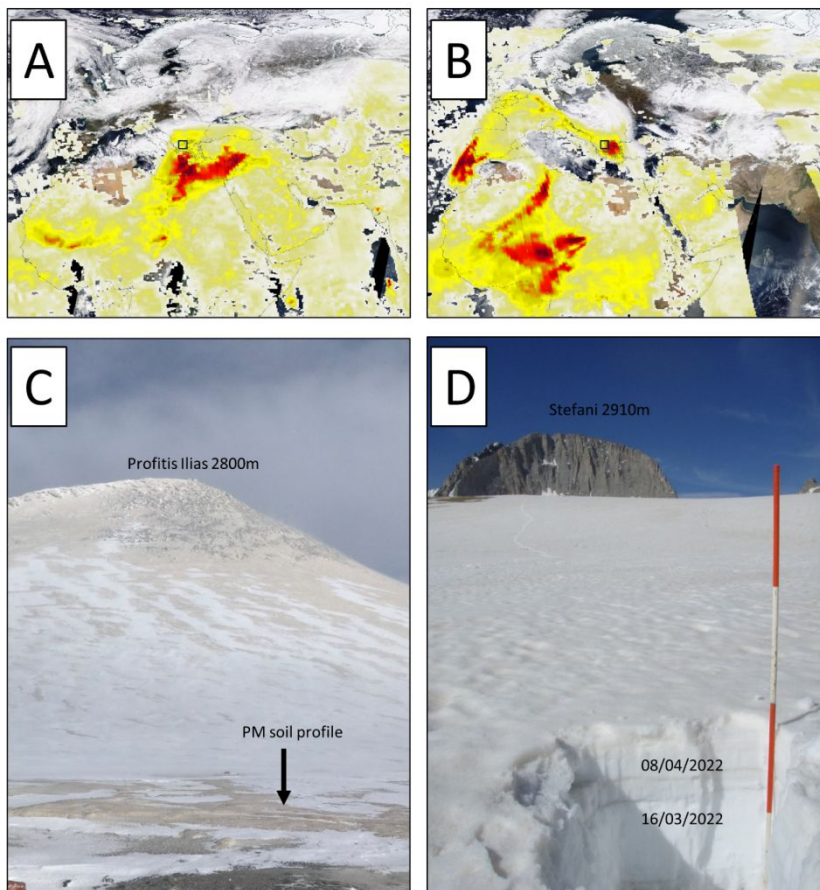
889 **Figure 2.** Conceptual diagram of the study, with the sampling sites and their morphological profiles,
 890 shown in Figure 1c as yellow lines, and with their respective textural characteristics that resulted
 891 from the grain size analysis. The respective heights of the rock cliffs (H_c) and talus slopes (H_t) are
 892 shown. The soil samples from the stratified scree clast free horizons in MK cirque are located behind
 893 the Little Ice Age moraine (left upper panel, photo, and diagram). The stratified scree slope under the
 894 rock wall of Stefani (2910m) in the TZ cirque, with the respective locations of the clast free soil
 895 samples (center panel, photo, and diagram) and the soil profile in the PM (right panel, photo, and
 896 diagram).



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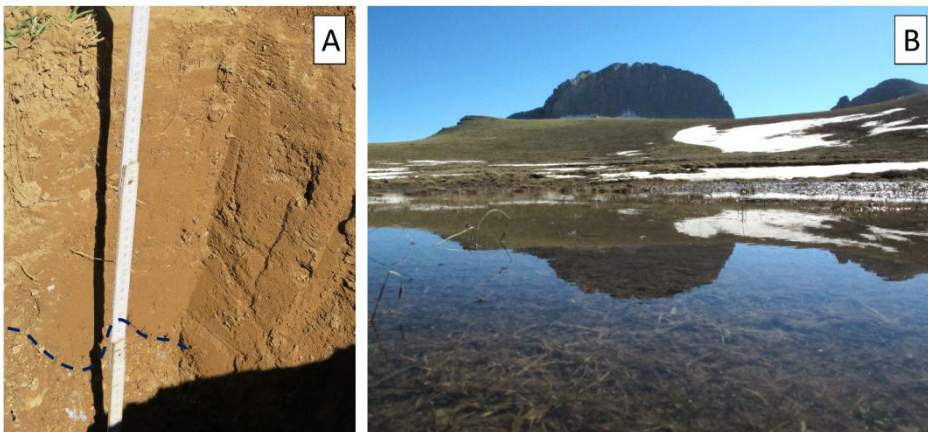
900 **Figure 3.** Synoptic maps and direct observations of two Sahara dust episodes on Mount Olympus
 901 alpine critical zone (black rectangle). (A). Aerosol Optical Depth (AOD) during March 22, 2018, and (B)
 902 and March 16, 2022, showing the trajectory of dust plume from Sahara Desert, and their impacts on
 903 the snowpack of the Plateau of Muses (C and D). The PM soil profile was excavated under the black
 904 arrow (C), whereas the snow pit (D) with two successive Sahara dust transport episodes in the spring
 905 of 2022, has also been excavated on top of the PM soil profile excavated pit. The NASA SUOMI/NNP
 906 Aerosol Optical Depth composition product was downloaded from the NASA EOSDIS Worldview
 907 platform (worldview.earthdata.nasa.gov).

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910 **Figure 4.** Evidence of soil disturbance on the Plateau of Muses under past and present-day climatic
911 conditions. (A) An irregular gravel layer (blue dashed line) between colluvial gravel and the overlying
912 soil resulting from cryoturbation. (B) Early summer season ongoing freeze of the soil surface layer
913 and subsequent waterlogging (photo taken on June 2012).

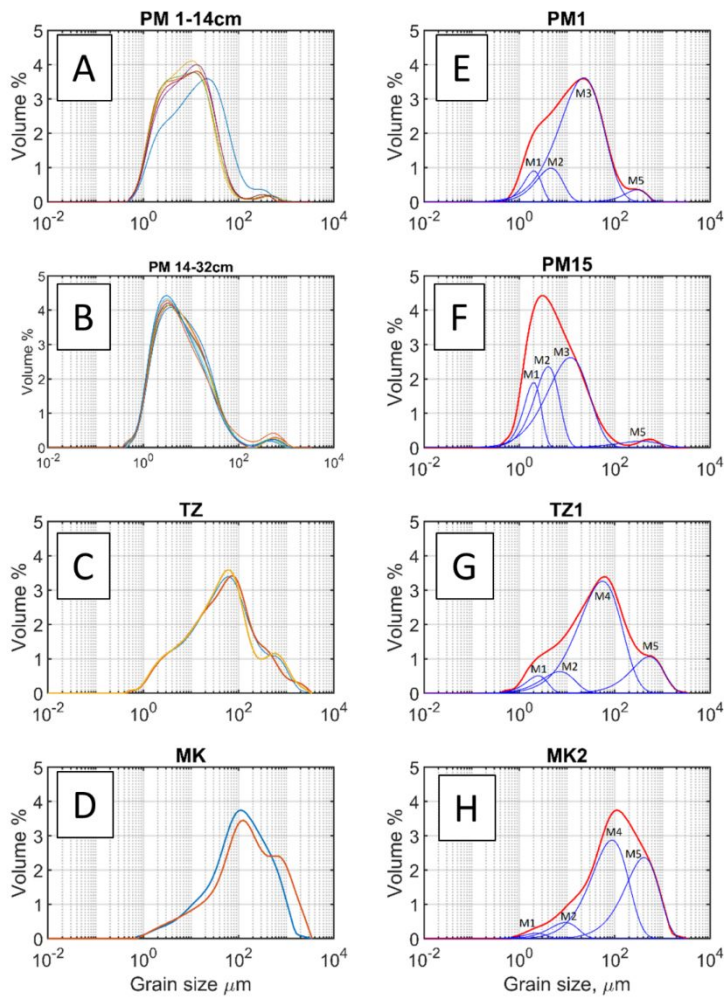


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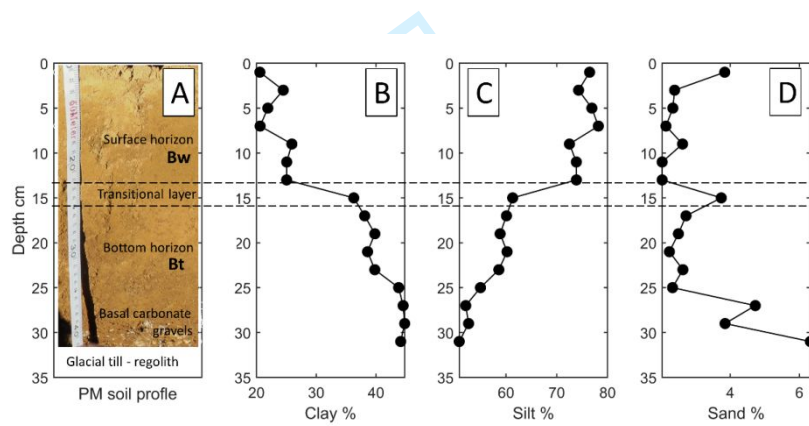
916 **Figure 5.** Cumulative grain size distributions of the soil samples from MK, TZ and PM Bw and Bt
 917 horizons (A, B, C and D). Surface sample PM1 (A, blue line) shows a distinct grain size distribution
 918 from the PM soil upper layer. Subplots E, F, G and H: Results of the CFLab fitting algorithm with the
 919 respective grain size distributions (GSD) and extracted grain size modes M1 to M5 (blue curves) of
 920 the soil samples PM1, PM15, TZ1 and MK2, represented as distinct sub-populations.

Commented [A25]: We clarify the use of separate grain size modes following the comment of Reviewer 2.



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922 **Figure 6.** The PM soil profile along with the depth variations main textural classes. A transition layer
 923 between 14 and 16 cm of depth marks a substantial decrease in clay and increase in silt contents and
 924 a change in the dry soil color. For clarity reasons, it is noted that the direct depth measurement of
 925 the PM soil begins at 10 cm along the measurement tape, explaining the discrepancy between the
 926 actual and the illustrated depth. The photo was taken one day after the profile excavation, when the
 927 upper part had partly dried out.

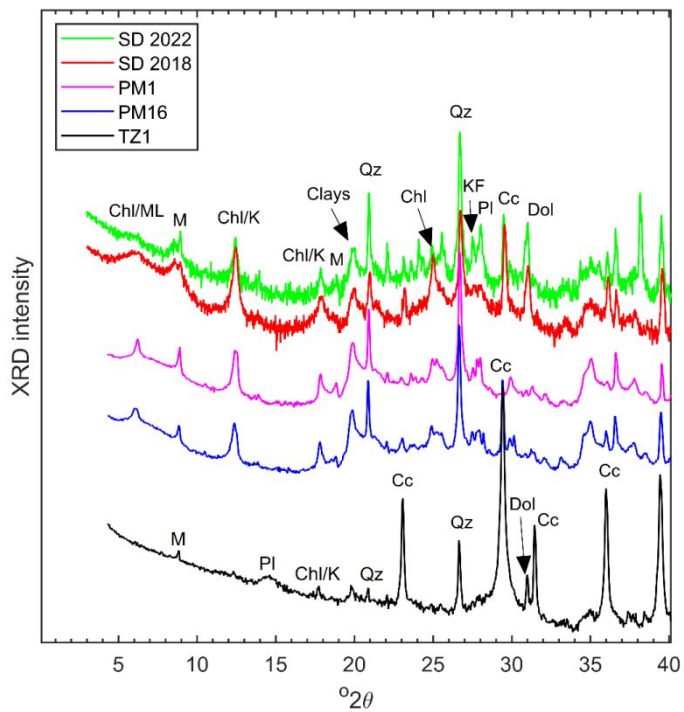


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931 **Figure 7.** X-ray diffraction patterns of soil samples PM1, PM15, TZ1 and MK2. Soils within the
 932 hillslope high energy scree deposits are composed primarily of calcite. In contrast PM soil samples
 933 contain quartz, clays, feldspars, and mica. (M: mica, Chl/M: Chlorite and mixed layer clays, likely
 934 smectite, Chl/K: chlorite and likely kaolinite, K: K feldspar, Qz: quartz, Pl: plagioclase feldspars, Cc:
 935 calcite, Dol: Dolomite).



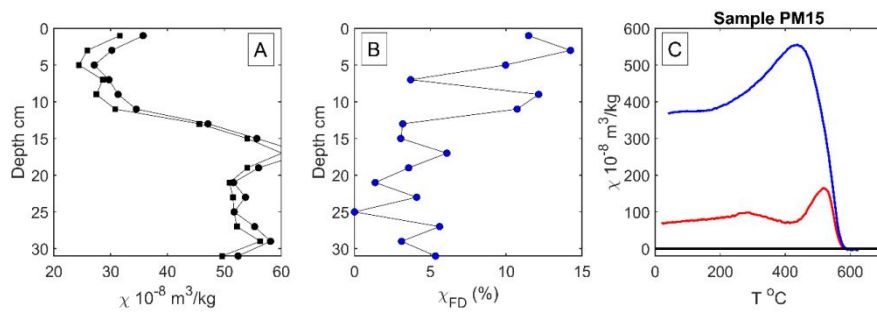
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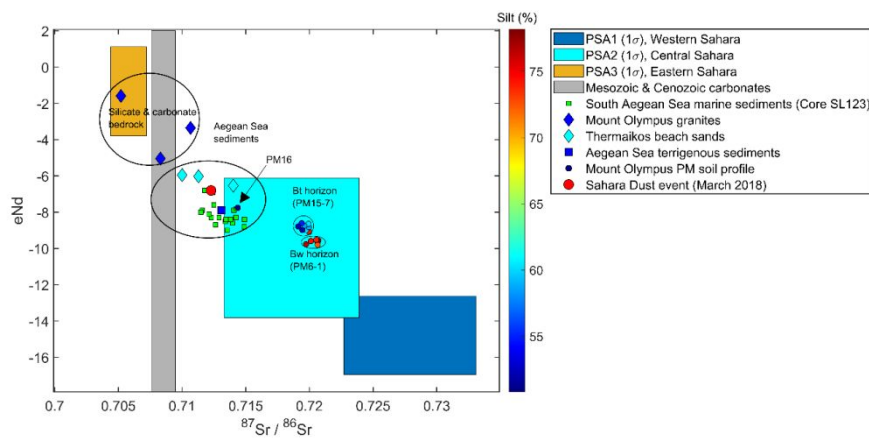
940 **Figure 8.** Depth variations of low and high frequency (A) and frequency dependent magnetic
 941 susceptibility (B) and (C) thermomagnetic analysis results of sample PM15 (red heating curve, blue
 942 cooling curve).



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946 **Figure 9.** Plot of $^{87}\text{Sr}/^{86}\text{Sr}$ against ϵ_{Nd} values of the PM soil samples with respective values of Mesozoic
 947 and Cretaceous carbonates (Franck et al., 2021), Mount Olympus granites (located on the continental
 948 west sides of the massif (Šarić et al., 2009, Castorina et al., 2020), Aegean Sea terrigenous and coastal
 949 sediments (Weldeab et al., 2002) and South Aegean marine sediments from Core SL 123 (Ehrmann et
 950 al., 2007), along with the three main North African dust source areas (PSA, Jewell et al., 2021). The
 951 isotopic enrichment trend for the PM soil samples towards crustal more radiogenic values occurs
 952 with a 25% increase in silt contents (colorbar) from the base to the surface of PM soil profile,
 953 suggesting the influence of external aeolian dust.

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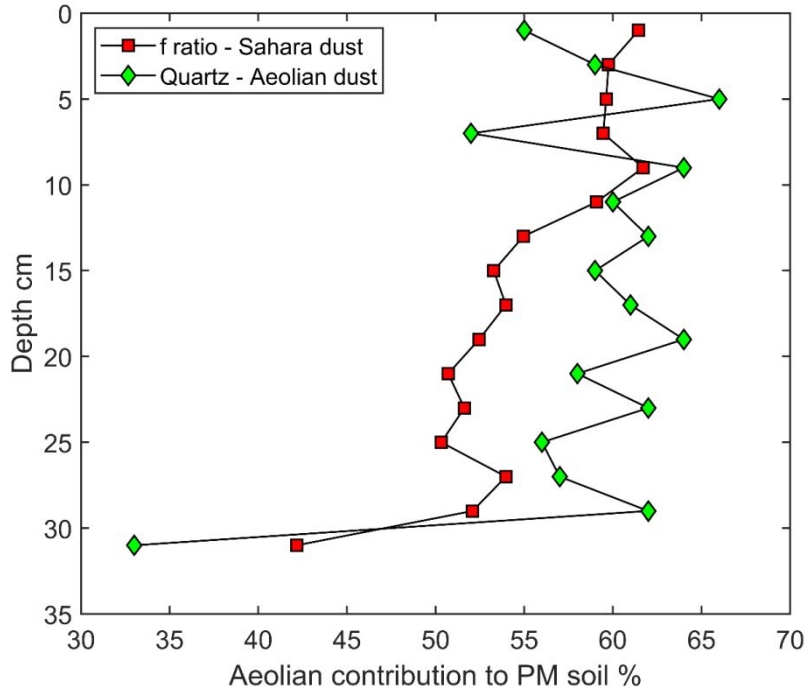
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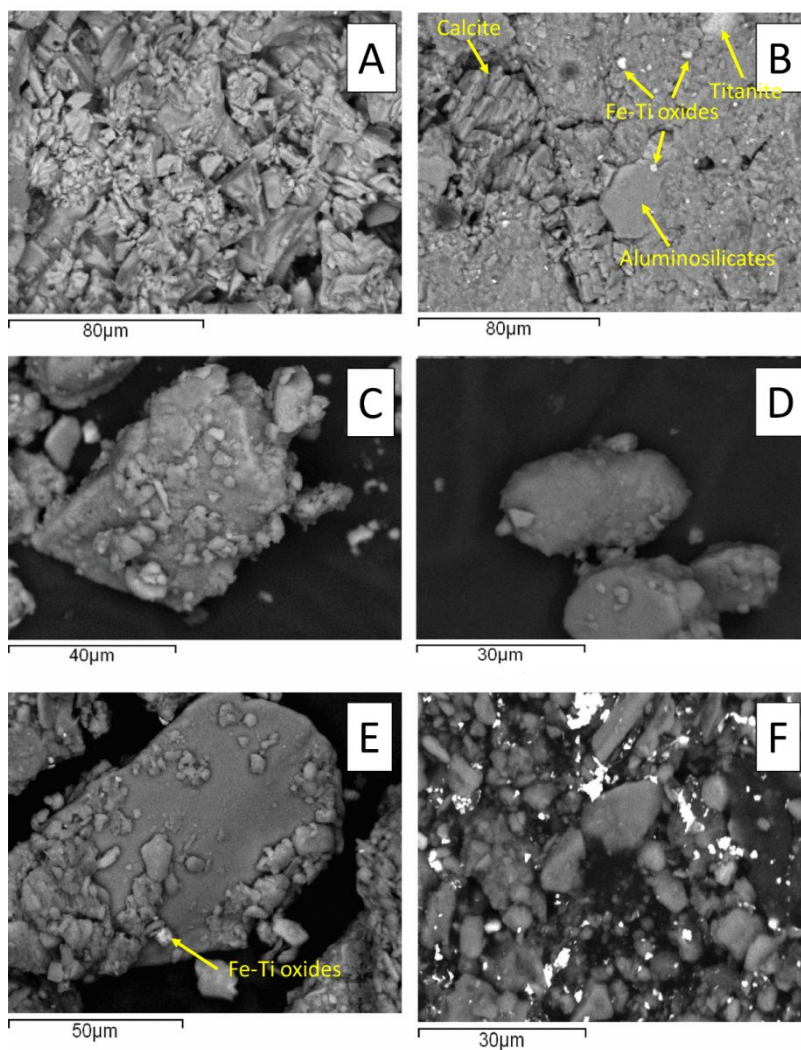
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961 **FIGURE 10.** Estimates of the relative contributions of aeolian dust accretion to PM soil as calculated
 962 by mineralogical, and isotopic proxies.



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971 **FIGURE 11.** SEM backscatter images from selected samples of PM loess profile. (A) Calcite grains
 972 from basal sample PM 16. (B) Mixed phase of aluminosilicates with calcite, titanomagnetite and
 973 titanite from basal sample PM 16. (C) K-feldspar. (D) Quartz grain with rounded edges as a result of
 974 long-range aeolian transport. (E) Surface sample PM1 aggregate of aluminosilicates and Fe-Ti oxides.
 975 (F) Quartz grains of variable shapes and grain sizes from sample PM3 along with Fe-Ti oxides.



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977 **Table 1.** Physical characteristics of the soil samples retrieved from the interbedded colluvial soils of
 978 the Megala Kazania (MK) and Throne of Zeus (TZ) scree slopes and from the alpine soil formed on the
 979 Plateau of Muses (PM).

Sample id	Depth (cm)	Munsell color (dry)	Clay (%) ($<2\mu\text{m}$)	Fine Silt (%) ($3.5\text{-}5\mu\text{m}$)	Silt (%) ($14\text{-}30\mu\text{m}$)	Fine sand (%) ($65\text{-}110\mu\text{m}$)	Coarse sand (%) ($300\text{-}800\mu\text{m}$)
			M1	M2	M3	M4	M5
MK1	30	5Y 6/1	1.7	6.6	0.0	39.8	51.9
MK2	250	5Y 6/1	0.0	11.4	0.0	47.5	41.1
TZ1	50	2.5Y 5/4	5.3	10.0	0.0	67.9	16.8
TZ2	150	10YR 3/4	6.0	25.2	0.0	43.5	25.3
TZ3	120	10YR 3/4	5.8	18.2	0.0	56.9	19.1
PM1	0-2	7.5YR 3/6	7.8	12.7	76.5	0.0	3.0
PM2	2-4	7.5YR 3/6	9.8	14.7	74.5	0.0	1.0
PM3	4-6	7.5YR 3/6	8.5	13.4	76.9	0.0	1.2
PM4	6-8	7.5YR 3/6	8.0	12.6	78.2	0.0	1.2
PM5	8-10	7.5YR 3/6	9.8	16.1	72.5	0.0	1.6
PM6	10-12	7.5YR 2/4	9.3	15.7	73.8	0.0	1.2
PM7	12-14	7.5YR 3/6	9.3	15.7	73.8	0.0	1.2
PM8	14-16	10YR 3/4	13.0	23.3	61.3	0.0	2.4
PM9	16-18	10YR 3/4	13.8	24.2	60.1	0.0	1.9
PM10	18-20	10YR 2/4	14.5	25.2	58.8	0.0	1.5
PM11	20-22	10YR 3/4	14.0	24.5	60.2	0.0	1.3
PM12	22-24	10YR 3/6	14.6	25.1	58.6	0.0	1.7
PM13	24-26	10YR 3/6	16.3	27.5	55.0	0.0	1.2

PM14	26-28	10YR 3/4	16.0	28.5	52.1	0.0	3.4
PM15	28-30	10YR 4/6	16.2	28.6	52.6	0.0	2.6
PM16	30-32	2.5YR 5/6	15.6	28.5	50.8	0.0	5.1

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982 **Table 2.** Weight percent (wt %) mineralogical semi quantitative composition of the PM soil, along
 983 with the clay mineralogy of surface and base samples PM1 and PM15.

Sample id	Qtz	Chl_CC	Plag	KF	Mica	Amph	Cc	Clay mineralogy (<2µm) Sm / Kaol / Chl / Ill
PM1	55	21	14	3	7	0	0	Smectite: 45%, Kaolinite: 35%, Chlorite: 10%, Illite:10%
PM2	59	26	6	1	8	0	0	
PM3	66	21	4	2	7	0	0	
PM4	52	31	7	3	7	0	0	
PM5	64	22	7	0	7	0	0	
PM6	60	24	7	2	7	0	0	
PM7	62	20	4	6	8	0	0	
PM8	59	24	7	3	7	0	0	
PM9	61	24	7	3	5	0	0	
PM10	64	22	8	1	5	0	0	
PM11	58	28	4	3	7	0	0	
PM12	62	22	7	2	7	0	0	
PM13	56	25	5	1	7	6	0	
PM14	57	28	2	7	6	0	0	
PM15	62	24	6	4	4	0	0	Smectite: 65%, Kaolinite: 25%, Chlorite: 5%, Illite:5%
PM16	33	12	1	2	4	0	48	

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986 **Table 3.** Radiogenic isotope results for the PM soil profile and the 2018 Sahara dust (SD) samples.

987 ϵ_{Nd} values were calculated

Sample id	$^{87}\text{Sr}/^{86}\text{Sr}$		$^{143}\text{Nd}/^{144}\text{Nd}$		ϵ_{Nd}
	$^{87}\text{Sr}/^{86}\text{Sr}$	std err (%)	$^{143}\text{Nd}/^{144}\text{Nd}$	std. err. (%)	
NBS 987 Sr standard	0.7102500	0.0008	-	-	
La Jolla Nd standard	-	-	0.5118500	0.0006	
PM1	0.7197322	0.0008	0.5121292	0.0004	-9.77
PM2	0.7201146	0.0026	0.5121383	0.0006	-9.60
PM3	0.7205816	0.0009	0.5121390	0.0006	-9.58
PM4	0.7207105	0.0010	0.5121399	0.0007	-9.56
PM5	0.7206530	0.0029	0.5121278	0.0009	-9.80
PM6	0.7205393	0.0007	0.5121419	0.0005	-9.52
PM7	0.7199828	0.0008	0.5121641	0.0005	-9.09
PM8	0.7196993	0.0009	0.5121731	0.0006	-8.91
PM9	0.7197594	0.0008	0.5121694	0.0005	-8.98
PM10	0.7199571	0.0007	0.5121775	0.0004	-8.82
PM11	0.7199244	0.0009	0.5121869	0.0006	-8.64
PM12	0.7196162	0.0008	0.5121820	0.0006	-8.74
PM13	0.7193930	0.0008	0.5121890	0.0006	-8.60
PM14	0.7194477	0.0009	0.5121694	0.0005	-8.98
PM15	0.7191162	0.0011	0.5121795	0.0004	-8.79
PM16	0.7143748	0.0016	0.5122328	0.0005	-7.75
SD 2018	0.7122721	0.0009	0.5122813	0.0006	-6.80

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