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

Michael Styllas, Christos Pennos, Aurel Persoiu, Athanasios Godelitsas, Lambrini Papadopoulou, Elina Aidona, Nikolaos Kantiranis, Mihai Ducea, Matthieu Ghilardi, François Demory

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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

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29

30

31 **Abstract**

32

33 Soil formation in Mediterranean periglacial landscapes remains poorly understood as the interplay  
34 between erosion and aeolian dust accretion in providing parent materials, and mineral weathering  
35 and pedogenesis, as dominant post depositional processes, depends on a variety of local and  
36 regional factors. Herein, we investigate the balance between erosion and aeolian dust accretion in  
37 the formation of an alpine soil profile along an erosional gradient in the periglacial zone of Mount  
38 Olympus in Greece. We applied a wide range of analytical methods to 23 samples, from a soil profile  
39 developed in a glaciokarstic plateau, from colluvial sediment horizons interbedded in postglacial  
40 scree slopes and from modern Sahara dust samples deposited on the snowpack. Colluvial sediment  
41 horizons exhibit high concentrations of calcite rich sand and represent the local erosion products.  
42 The soil B horizon developed on a glaciokarstic plateau contains high amounts of fine earth and is  
43 rich in quartz, mica, plagioclase, clays, and Fe-Ti oxides. Based on its physical and textural  
44 characteristics the soil profile is partitioned in a surficial weathered Bw and a lower illuvial Bt horizon  
45 that overlies the local regolith composed of fragmented glacial till and slope wash deposits.  
46 Radiogenic isotope systematics, textural and mineralogical analysis show that the contribution of  
47 Sahara and locally sourced dust to the development of the soil B horizon ranges between 50 and  
48 65%. Cryoturbation results in fine earth translocation from Bw to the Bt horizon, whereas weak  
49 pedogenetic modifications of aeolian and bedrock-derived minerals result in magnetic mineral  
50 weathering and secondary clay formation. Our findings reveal that, aeolian dust accretion is the  
51 dominant process in providing alpine soil parent material and that cryoturbation, weak pedogenesis,  
52 and clay mineral alteration occur within the Mediterranean periglacial zone of Mount Olympus.

53

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

56

57

58        **1. INTRODUCTION**

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

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

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

87 Most soils formed in the Mediterranean basin display a distinguishable red color (terra rossa)  
88 that derives from high concentrations of ultra-fine pedogenic iron oxides, mainly hematite (Yaalon,  
89 1997; Durn et al., 1999). Terra rossa soils receive significant aeolian dust additions from Sahara and  
90 Sahel regions (Yaalon, 1997; Durn, 2003; Stuut et al., 2009). In the Mediterranean alpine hinterland,  
91 thin drapes of Sahara-dust-rich soils are found on plateaus, glacial moraines, and outwash plains  
92 (e.g., Rellini et al., 2009), whereas aeolian dust accretion in terra rossa soils can also originate from a  
93 wide range of alluvial deposits, such as sand dunes, desiccated alluvial planes, and Quaternary loess  
94 (Amit et al., 2020; Lehmkuhl et al., 2020). Most of the Mediterranean mountains are built up by  
95 carbonate rocks, hence the aeolian input to alpine soil formation occurs in parallel with colluvial  
96 deposition of carbonate erosion and dissolution products that form a characteristic insoluble residue  
97 incorporated in the soil sequences (Durn 2003; Varga et al., 2016; Kirsten and Heinrich, 2022).

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

110 the sedimentological, mineralogical, and radiogenic isotope compositions through the application of  
111  $^{86}\text{Sr}/^{87}\text{Sr}$  and  $^{143}\text{Nd}/^{144}\text{Nd}$  ratios between the soil samples and Sahara dust samples collected from the  
112 snowpack. We finally measured the magnetic properties of the soil samples and clay mineralogy of  
113 bottom and topsoil layers, to assess the potential for weathering of clay minerals and iron oxides  
114 within Mount Olympus periglacial zone. Understanding the sources of parent materials and soil  
115 formation processes between contrasting geomorphological settings is a fundamental step towards  
116 defining the postglacial paleo-environmental history of Mount Olympus alpine landscapes that  
117 followed pronounced shifts of the regional climate.

118

## 119 **2 BACKGROUND**

### 120 **2.1 Mount Olympus glacial history**

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

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

142

## 143 **2.2 Climate**

144 The contemporary maritime conditions and the steep relief of Mount Olympus result in intense  
145 precipitation and temperature altitudinal gradients, with the highest peaks constituting an  
146 orographic and climatic barrier between the eastern (marine) and western (continental) sides (Figure  
147 1b, Styllas and Kaskaoutis, 2018). The climate in the coastal zone is typically Mediterranean, whereas  
148 at higher elevations (1000–2200 m), the climate attains sub-Mediterranean characteristics with  
149 average annual precipitation of 1300 mm (Styllas et al., 2016). In the alpine zone above the tree line  
150 (2400 m), the climate is characterized by temperate conditions with annual precipitation above 2000  
151 mm and average annual temperatures between 0 and 1.5 °C (Styllas et al., 2016). The periglacial  
152 activity in the Mount Olympus alpine zone is likely still active today, as it is situated just above the  
153 lower limit of the regional permafrost zone (2700m) of the southern Balkan peninsular (Dobiński,  
154 2005).

155

## 156 **2.3 The Plateau of Muses**

157 The Plateau of Muses (PM) is a planar depositional surface located at an elevation of 2600 m with a  
158 surface area of 1 km<sup>2</sup>. It resembles a typical alpine meadow, partly covered by alpine grass  
159 vegetation that shares similar characteristics with plateaus found in the high Balkan Mountains and

160 the European Alps. The PM is bounded to the south by the TZ cirque lateral moraine ridge and by  
161 several gentle-sloping glacially eroded peaks along its northern, eastern, and western margins (Figure  
162 1C). The formation of the plateau has resulted from the combined action of glacial scouring and  
163 carbonate bedrock dissolution. Its low relief in combination with the circular shape suggest a doline  
164 type karstic depression that is filled with glacial till, overlain by colluvial sediments (slope wash)  
165 transported from the adjacent slopes. The surface layer of the PM sedimentary sequence comprises  
166 a developed soil sequence with variable thickness (30–50 cm) that overlies a layer of outwash sand  
167 and fine gravels and/or fragmented till boulders, and exhibits brown-red to yellow color hues, which  
168 in the Munsell color scale range between 7.5 and 10 YR (Table 1). Alternating patches of alpine grass  
169 vegetation and hummocky soil pans in the center of the plateau are indicative of periglacial activity  
170 and cryoturbation. Other periglacial features such as solifluction-terraced stripes below the bare  
171 bedrock of the surrounding summits are tentatively considered to have formed during the Late  
172 Holocene cold stages, during the observed expansion of small glaciers in the MK cirque.

173

### 174 **3 MATERIALS AND METHODS**

#### 175 **3.1 Erosional products and alpine soil sampling**

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



186 similar textural, mineralogical, and geochemical characteristics with the PM soil basal horizon, which  
187 lies on a layer of outwash sand and gravels. Luckily, we were able to retrieve the samples from two  
188 distinct interbedded clast free sediment layers within the TZ scree slope after a torrential rainfall  
189 event that opened a deep erosional trench in the scree below the rockwall and reached the basal till  
190 layer (Figure 2). The scree slope in the MK is regularly eroded and scoured from a perennial snowfield  
191 that is retreating by the end of the summer season, and this made the sampling of distinct soil-  
192 sediment horizons straightforward. We manually excavated only one pit for high resolution soil  
193 sampling and considered that due to the very small surface area of the surficial soil apron within the  
194 PM catchment (0.06 km<sup>2</sup>), the specific profile is representative of the PM soil development. We  
195 selected a location in the center of a circular soil-sediment pan that was free of vegetation, surface  
196 carbonate fragments (Figure 2). After sampling, the pit was closed and refilled with the excavated  
197 material in accordance with Mount Olympus National Park directions. In locations with long lasting  
198 snowpack, we observed a humic A horizon, but since these locations host several endemic flower  
199 species, the Management Unit of Mount Olympus National Park, did not grant permission to  
200 excavate a soil pit in these sensitive sites. The PM soil samples were additionally subjected to  
201 microscopic and radiogenic isotope analyses and magnetic measurements to investigate the  
202 potential chemical alterations processes during PM soil development. Mineralogical and radiogenic  
203 isotope analyses were also performed in two (n = 2) samples of aeolian dust that were deposited on  
204 the PM snowpack during the spring seasons of 2018 and 2022. The long-range aeolian dust transport  
205 episodes occurred on March 22–24, 2018, and March 16–18, 2022. The synoptic conditions of these  
206 distinct episodes show that the dust emissions traveled to Mount Olympus from the Sahara Desert  
207 and left an orange hue on the snowpack, which later in the spring season formed distinct layers in  
208 the snowpack (Figure 3). We therefore consider the samples collected from the PM snowpack as  
209 representative of Sahara dust accretion in Mount Olympus alpine soils.

210

### 211 **3.2 Grain size analyses**

212 The soil samples were transported to the lab, wet sieved through a 3.5-mm sieve, and treated with  
213 30% hydrogen peroxide (H<sub>2</sub>O<sub>2</sub>) at 70 °C for 12 h to remove organic matter. The H<sub>2</sub>O<sub>2</sub> treatment was  
214 repeated three times until the samples were completely bleached and all organic matter was  
215 degraded. The samples were washed with distilled water and analyzed with a Mastersizer 3000 laser  
216 diffraction particle-size analyzer to define the bulk grain size distributions of the sand, silt, and clay  
217 fractions. The samples were run through the automated dispersion unit and sodium  
218 hexametaphosphate solution (Calgon) was added as dispersion factor. Statistical analyses of the grain  
219 size distributions and derivation of the clay, silt, and sand fractions were realized with MATLAB Curve  
220 Fitting Lab (CFLab), which performs curve fitting on sediment grain size distributions using the  
221 Weibull probability distribution function (Wu et al., 2020).

222

### 223 **3.3 Mineralogy**

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

232

### 233 **3.4 Petrographic, magnetic, and isotopic analyses**

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

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

242 We additionally explored the existence of ferromagnetic components and the potential for  
243 secondary iron oxides formation in the PM soil profile through magnetic susceptibility  
244 measurements. The discrete samples were packed in cubical plastic boxes ( $2 \times 2 \times 2$  cm) and weighed  
245 before the measurements. Volume-specific magnetic susceptibility measurements were performed  
246 using both a Bartington dual MS2B sensor at low and high frequencies of 0.465 and 4.65 kHz. The  
247 results are expressed as mass-specific magnetic susceptibility ( $\chi$ ;  $10^{-8}$  m<sup>3</sup>/kg). During the measuring  
248 procedure, every sample was measured at least three times and the average value was assigned as  
249 the final measurement. For each sample, two air measurements were performed before and after  
250 sample measurement. The frequency-dependent susceptibility ( $\chi_{FD}$ ; %) was calculated according to  
251 Dearing et al. (1996):

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

252  
253 where  $\chi_{LF}$ ,  $\chi_{HF}$ , are the magnetic susceptibility at low and high frequency, respectively. Samples PM16  
254 and PM15, which were considered as more representative of the PM soil regolith boundary, were  
255 additionally subjected to thermomagnetic analysis to define the origin of the ferromagnetic particles  
256 at the base of the PM soil. Measurements of continuous thermomagnetic curves (K–T curves) at low  
257 and high temperature were realized with the furnace CS3 of the AGICO MFK1-FA susceptibilimeter.

258 The potential sources of the PM soil and aeolian dust were evaluated through their Sr and Nd  
259 isotopic ratios. Isotopic measurements were performed at the University of Arizona TIMS laboratory  
260 following the procedure in Conroy et al. (2013) on soil samples. Samples were not spiked and  
261 dissolved in mixtures of ultrapure Hf-HNO<sub>3</sub> acid. Elemental separation of dissolved samples was  
262 carried out in chromatographic columns via HCl elution in a clean laboratory environment.  
263 Conventional cation columns filled with AG50W-X4 resin were used for Sr and REEs separation and

264 anion columns with LN Spec resin for Nd separation following Ducea et al. (2020). Sr cuts were  
265 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  
266 3) were measured on a VG Sector 54 thermal ionization mass spectrometer (TIMS) fitted with  
267 adjustable 1011  $\Omega$  Faraday collectors and Daly photomultipliers. NBS SRM 987 Sr standard and La  
268 Jolla Nd standard were analyzed during the samples run to ensure the performance of the  
269 instrument and to perform some minor correction on the final reported ratios.

270

### 271 **3.5 Erosional potential and aeolian dust accretion proxies**

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

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

288 We selected the  $\epsilon_{\text{Nd}}$  ratio as a second independent proxy particularly of Sahara dust accretion  
289 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

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

297

$$298 \quad f = \frac{(\epsilon_{Nd} \text{ PM soil} - \epsilon_{Nd} \text{ bedrock})}{(\epsilon_{Nd} \text{ Sahara dust} - \epsilon_{Nd} \text{ bedrock})} \quad (2)$$

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

303

## 304 **4 RESULTS**

### 305 **4.1 Alpine soil formation across a hillslope energy gradient**

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

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

324

#### 325 **4.2 Grain size variation**

326 The interactions between slope processes, colluvial sediment transport, and aeolian sub additions  
327 result in polymodal grain size distributions that display different shapes among MK, TZ, and PM soils.  
328 Five grain size modes (M1 to M5) were mathematically derived from the application of the CFLab  
329 curve-fitting algorithm. Fitting degrees were >99% and fitting residuals were <0.1%, indicating  
330 excellent fits for the raw grain size distribution curves (Figure 5 A, B, C, and D). The fine-earth (clay  
331 and silt) fractions resemble grain size modes M1 and M2 with respective mean grain sizes of  $\sim 2$  and  
332  $\sim 4$   $\mu\text{m}$  and M3 with mean grain sizes between 10 and 30  $\mu\text{m}$ . The sand fraction is composed of two  
333 modal sub-populations: a fine-sand-grain size mode (M4, mean grain size  $\sim 80$   $\mu\text{m}$ ) and a coarse  
334 sand-grain size mode (M5, mean grain size 440  $\mu\text{m}$ ) (Figure 5 E, F, G, and H; Table 1). The production  
335 of coarse sand is transported to the respective interbedded sediment horizons by rockfall activity and  
336 colluvial processes, or in the case of the low-sloping PM, through slope wash. The fine sand (M4)  
337 subpopulation was not traced in the PM soil samples, and this can be linked to either selective  
338 entrainment of M4 or to distortion of the MK and TZ grain size curves and truncation of the coarser  
339 modes (Garzanti et al., 2009).

340 In addition to the distinct color variations, the contrasting slope-energy distribution between the  
341 MK, TZ scree slopes and the PM depositional environments also defines their textural compositions.  
342 Sediment horizons developed on the surface of the MK scree slope contain higher amounts of sand  
343 (~90%) and lower amounts of silt and clay (~10%) compared with their TZ counterparts (~75% and  
344 ~25%), implying that the dominance of sand in the sediment horizons of the scree slopes derives  
345 from freeze-thaw and colluvial activity. The coarse-sand content (M5) of the PM soil basal layer is 6%  
346 but is lower within the solum (2%–3%), suggesting either reduced periglacial activity and/or low  
347 transport capacity of erosional products from the catchment through slope wash processes during  
348 the PM soil formation (Table 1; Figure 2, lower graphs).

349 The grain size distribution curves of the PM soil present a significant change in shape between  
350 soil depths of 14 and 16 cm, which is characterized by a 15% reduction of the clay and very-fine-silt  
351 fractions (M1 and M2) and by a similar increase of silt contents (Figure 5A and B). This sharp textural  
352 differentiation was not supported from field observations, where the solum appeared homogenous  
353 without distinct pedogenetic horizons and without any visual evidence of an erosional layer (Figure  
354 6A), but it is supported by changes in the soil color. The samples above a soil depth of 14–16 cm  
355 exhibit red to brown hues (7.5 YR), whereas the samples below this layer have more yellow-red (10  
356 YR) hues (Table 1). We also observed clay coatings in sparse secondary carbonates (calcretes) along  
357 the lower part of the PM soil profile, which we interpret as evidence of soil mixing and downward  
358 translocation of dissolved Ca and secondary calcite precipitation at the base of the soil profile. Based  
359 on these observations, we partitioned the PM soil profile in two horizons: an upper Bw horizon  
360 between 0 and 14 cm with red to brown hues, low clay (~25%), and high silt (~75%) contents, and a  
361 lower illuvial Bt horizon between 14 and 32 cm with higher (~40%) clay contents, a yellow-red hue,  
362 and smaller amounts (~5%) of sand compared with the overlying Bw horizon (Figure 6).

363

364

### 365 4.3 Soil and aeolian dust mineralogy

366 XRD analysis of the bulk samples reveals a mineralogy that substantially differs between the MK,  
367 TZ, and PM soils and, like the soil colorization and textural variations, follows the erosional slope  
368 gradient. The most dominant mineral phase in the clast-free material of the MK and TZ soils is calcite.  
369 Other minerals identified include dolomite along with quartz and micas. Conversely, the bulk  
370 mineralogical composition of PM soil exhibits a richer matrix of minerals that includes quartz, chlorite  
371 and mixed layer clays, mica, potassium feldspars, and plagioclase (Figure 7). Calcite is dominant  
372 (~50%) only in basal sample PM16 (Figure 6; Table 2). Quartz, clays, and mica are the most dominant  
373 mineral phases in the PM soil (~80%) with low values in basal sample PM16, whereas plagioclase, K-  
374 feldspar, and mica represent the remaining 20% (Table 2). Semi-quantitative analysis of the clay  
375 mineralogy of two samples retrieved from the surface of the Bw horizon and the base of the Bt  
376 horizon (samples PM1 and PM15) revealed high concentrations of smectite and kaolinite (80%) and  
377 low contents of chlorite and illite. Surface sample PM1 contains 45% smectite and 35% kaolinite,  
378 whereas basal sample PM15 has higher smectite (65%) and lower kaolinite (25%) contents (Table 2).

379 From the comparison of the XRD spectra (Figure 7), it is obvious that the bulk mineralogy of the  
380 PM soil matches that of the Sahara dust samples. Both Sahara dust samples show the presence of  
381 clay minerals, quartz, mica, calcite, plagioclase, K-feldspar, and dolomite. The detected mineral  
382 phases are typical of Saharan dust deposited in Europe during both dry- and wet-deposition (red  
383 rains) events (Scheuvens et al., 2013). Additionally, recent studies of Saharan dust wet deposition in  
384 the Iberian Peninsula also indicated the presence of Fe-Ti oxides, such as goethite and hematite, and  
385 of Ti oxides, such as rutile (Rodriguez-Navarro et al., 2018), but these were not depicted from our  
386 XRD analyses. Despite their overall XRD spectral similarity, a pronounced difference between the  
387 contemporary Sahara dust and PM soil samples is the presence of calcite and dolomite in the dust  
388 samples and their near absence from the PM soil profile (Figure 7). The smooth and low intensity  
389 peaks for calcite and dolomite at 29.43 and 30.7 2 $\theta$  in surface sample PM1 indicate the partial



390 removal of calcite, whereas similarly subdued peaks in basal sample PM16 denote near complete  
391 decalcification of the solum (Figure 7).

392

#### 393 **4.4 Magnetic susceptibility of PM soil**

394 The magnetic susceptibilities of the PM soil bulk samples were measured to provide insight into the  
395 ferromagnetic components of the PM soil and their potential alterations. Overall, the low-frequency  
396 magnetic susceptibility ( $\chi_{lf}$ ) is higher in the lower Bt horizon, with average values for samples PM8–  
397 PM16 of  $55 \times 10^{-8} \text{ m}^3 \text{ kg}^{-1}$ , and lower  $\chi_{lf}$  values in the Bw horizon, with average values for samples  
398 PM1–PM7 of  $36 \times 10^{-8} \text{ m}^3 \text{ kg}^{-1}$  (Figure 8A). Similar value ranges were measured for the high-  
399 frequency magnetic susceptibility ( $\chi_{hf}$ ). The estimated values of frequency-dependent ( $\chi_{FD}$ )  
400 susceptibility presenting a wide range of values ranging between 0% (sample PM13) and 14%, with  
401 significantly higher values in the Bw horizon (Figure 8B). According to Dearing (1999), high  $\chi_{FD}$  values  
402 (>10%) are indicative of the presence of superparamagnetic Fe oxide nanoparticles (< 0.05  $\mu\text{m}$ ),  
403 suggesting a higher amount of fine ferrimagnetic grains in the surface horizon Bw, which potentially  
404 can be of detrital (aeolian and/ eroded bedrock) origin.

405 The mineral phases responsible for the magnetic enhancement of the Bt horizon were deduced  
406 from high-temperature magnetic susceptibility measurements performed during a single heating–  
407 cooling cycle to 700 °C (Figure 8C). We estimated the Curie temperature ( $T_c$ ) of samples PM16 and  
408 PM15 to examine the potential existence of superparamagnetic ultrafine particles in the base of the  
409 PM soil profile, which is in contact with the regolith. The thermomagnetic analysis of sample PM16  
410 failed completely, likely due to its high calcite content and absence of magnetic phases. On the other  
411 hand, sample PM15 resembling the soil-regolith lower boundary, revealed a uniform  $\chi$ -T behavior  
412 that is indicative of the dominance of two magnetic phases (Figure 8C) – one with  $T_c$ , or  
413 transformation temperature, between 260–320 °C, probably maghemite, and a second one around  
414 600 °C, which is typical for oxidized magnetite (Jordanova et al., 2022). Since the nano-sized

415 pedogenic magnetite is unstable upon heating (Dunlop and Özdemir, 1997), the identified oxidized  
416 magnetite suggests that weak pedogenetic production of ferromagnetic components occurs in the  
417 base of PM soil profile.

418

419

#### 420 **4.5 Radiogenic isotopes**

421 More information on the provenance of the PM-soil-forming material was derived from the  
422 radiogenic isotope analysis of the soil samples and of the 2018 Sahara dust sample. The  $^{87}\text{Sr}/^{86}\text{Sr}$   
423 values of PM soil samples range from 0.71437 to 0.72071 and the  $\epsilon_{\text{Nd}}$  values from -7.75 to -9.80  
424 (Table 3). Overall, the PM soil  $^{87}\text{Sr}/^{86}\text{Sr} - \epsilon_{\text{Nd}}$  cluster together apart from sample PM16, which has the  
425 lowest value of the PM soil  $^{87}\text{Sr}/^{86}\text{Sr}$  ratio (Figure 9). The analyzed Sahara dust sample exhibits  
426  $^{87}\text{Sr}/^{86}\text{Sr}$  value of 0.71272 that falls within the lower range of North African dust sources between  
427  $\sim 0.71200$  and  $0.74000$  (Erel and Torrent, 2010; Grousset and Biscaye, 2005). The Sr isotopic ratio of  
428 the Sahara dust sample shows potential mixing with rainwater and local sea salt aerosols during the  
429 March 2018 wet deposition event but also with other European aerosol sources, which is validated  
430 by the fact that the dust plume of the March 2018 travelled over Europe before it reached Mount  
431 Olympus (Figure 3A). The Sahara dust sample has an  $\epsilon_{\text{Nd}}$  value of  $-6.80$ . Plotting the  $^{87}\text{Sr}/^{86}\text{Sr}$  and  $\epsilon_{\text{Nd}}$   
432 measurements against literature values from terrigenous, coastal, and marine sediments from the  
433 Aegean Sea region (Weldeab et al., 2002) reveals an isotopic similarity between the Sahara dust and  
434 of sample PM16 with these sediments (Figure 9). A reasonable interpretation of this observation  
435 comes from the fact that basal sample PM16 resembles more the soil regolith and plots close to the  
436 contemporary and Holocene values of Aegean Sea terrestrial, coastal and marine sediments. The  
437  $^{87}\text{Sr}/^{86}\text{Sr}$  values representing the PM soil regolith show similar values with the Aegean Sea terrestrial  
438 and marine sediments and likely represent a mix of Sahara dust with Mesozoic and Cenozoic bedrock  
439 carbonates, which are overall characterized by low  $^{87}\text{Sr}/^{86}\text{Sr}$  values of  $<0.70800$  (Capo et al., 1998;  
440 Frank et al., 2021). The two subclusters of PM soil samples have more radiogenic values compared

441 with those of the rest of the samples and clearly correspond to Bw and Bt horizons. The increasing  
442 silt contents towards the surface of the PM soil profile (Figure 6, Table 1) occur with  $^{87}\text{Sr}/^{86}\text{Sr}$  and  $\epsilon_{\text{Nd}}$   
443 values towards more crustal values (color variation in Figure 9) that are representative of the central  
444 Sahara province. Therefore, the increases in the silt fraction within the PM soil profile can be directly  
445 linked to increases in Sahara dust accretion. This is further supported by range of the silt fraction  
446 mean grain size between 14 and 30  $\mu\text{m}$  (Table 1), which is similar to those for modern Sahara dust  
447 deposits from Crete, which range 4–8  $\mu\text{m}$  and 16–30  $\mu\text{m}$  (Mattson and Niéhlen, 1996; Goudie and  
448 Middleton, 2001). In terms of Sahara dust provenance fingerprinting, the  $^{87}\text{Sr}/^{86}\text{Sr}$  and  $\epsilon_{\text{Nd}}$  values of  
449 the PM soil samples fall within the range ( $1\sigma$ ) of the central North African dust source area, which  
450 broadly involves the Bodele depression (PSA2; Jewell et al., 2021).

451

## 452 **5 DISCUSSION**

### 453 **5.1 PM soil parent material**

454 The mineralogical (XRD) analyses, show that calcite is the dominant mineral phase of MK and TZ  
455 interbedded sandy sediment and PM basal layers (Figure 7, lower XRD diagrams TZ01 and Table 2  
456 sample PM 16), which in the periglacial environment of Mount Olympus is expected to dissolve  
457 slowly (e.g., Gaillardet et al., 2019) and produce an insoluble residue that comprises the PM soil  
458 parent material. MacLeod (1980) analyzed the mineral composition of the insoluble residue of  
459 carbonates from western Greece and defined a mineralogical suite of quartz, kaolinite, and mica  
460 (illite). Kantiranis (2001) studied the carbonate rocks of northwestern Greece and found insoluble  
461 residue ~1wt.% consisting mainly of micas, quartz, hematite, chlorite, feldspars, and amphibole,  
462 whereas the insoluble residue of carbonate basement rocks from Crete also resembles ~1 wt.% of  
463 the whole rock samples and is composed of a sandy loam matrix rich in quartz, plagioclase (albite),  
464 and mica (illite) (Kirsten and Heinrich, 2022). Thus, the dissolution of the local carbonate parent  
465 material within the interbedded sediment layers and in the basal layer of PM soil, can release very

466 small quantities of bedrock-derived impurities such as quartz, plagioclase, illite, and kaolinite that are  
467 incorporated in the solum, but cannot explain the ~30cm thick PM soil mantle and ~60cm thickness  
468 of the layers interbedded in the scree slopes.

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

491

## 492 **5.2 Relative contributions of aeolian dust inputs**

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

509 The  $\epsilon_{\text{Nd}}$ -based Sahara dust contributions to the PM soil varies between  $\sim 35\%$  and  $\sim 50\%$   
510 (except that of basal sample PM16) (Figure 10). Conversely, the quartz-derived aeolian dust  
511 contribution ranges between  $\sim 45\%$  and  $65\%$ , shows a relatively small variation with depth and an  
512 abrupt increase ( $\sim 25\%$ ) from sample PM16 to PM15 (Figure 10). The basal sample PM16 exhibits the  
513 lowest contributions of quartz concentration,  $f$  ratio values (Figure 10) and silt concentrations (Table  
514 1, Figure 6) and is considered an outlier representing the regolith-PM soil mix, which agrees with its  
515 distinct color and lowest magnetic susceptibility values. The preservation of quartz in the PM soil

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

525         Based on the above, we tentatively attribute the  $\sim 15\%$  difference between the  $\epsilon_{Nd}$ -based  
526 estimates and the quartz-based estimates to accretion of quartz-rich dust from local sources during  
527 the formation of the Bt horizon, considering that the contribution of bedrock derived quartz from  
528 the insoluble residue is  $\sim 1\%$ . From the  $\epsilon_{Nd}$ -based contributions, we estimate that the Sahara dust  
529 accretion to PM soil is between  $\sim 35\%$  to  $50\%$ , whereas local sources can potentially accrete another  
530  $\sim 15\%$ . Our estimated aeolian dust accretion  $\sim 65\%$  is similar to the one in the North Calcareous Alps,  
531 where the local contribution of dust from the periglacial zone of the North Calcareous and Austrian  
532 silicate Alps is significant (Küfmann 2008), but our estimated Sahara dust contribution in the PM soil  
533 is higher than its respective average contribution ( $20 - 30\%$ , Varga et al., 2016) in interglacial soils of  
534 the Carpathian Basin. We attribute this difference to the closer proximity of Mount Olympus to  
535 Sahara Desert than the Carpathian basin. Given that these values are conservative estimates, the  
536 aeolian contribution may potentially be higher as, in our calculations, we have not included aeolian  
537 transported micas, feldspars, and clays that are integral parts of Sahara dust samples deposited on  
538 the snowpack. We thus suggest that the aeolian dust accretion comprises a minimum of  $\sim 65\%$  of the  
539 PM soil parent material and that carbonate bedrock erosion, and pedogenetic production of detrital  
540 clays can potentially contribute another  $\sim 35\%$  to the development of PM soil.

541

### 542        **5.3 Pedogenetic alterations**

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

555        Despite the absence of color difference and of distinct layers in the PM soil, the higher magnetic  
556        susceptibility values of the Bt compared to the Bw horizon, can result from the enrichment of  
557        ferromagnetic minerals during *in situ* weathering of translocated detrital fine earth particles through  
558        pedogenesis (Maher, 2011). However, the overall low values of the frequency dependent magnetic  
559        susceptibility ( $\chi_{FD}<10$ ), point to weak pedogenetic alteration of soil (Dearing et al., 1996), which in the  
560        base of PM soil profile occurs through the oxidation of ultrafine (titano)magnetite to maghemite  
561        (Section 4.4). The SEM-EDS analyses show the presence of ultrafine Fe-Ti grains throughout the  
562        solum (Figure 11B) apart from basal sample PM16 (Figure 11A), which is representative of the PM  
563        soil regolith. This is further supported from the EDS chemical composition of the calcite grains in  
564        basal sample PM 16 that have TiO<sub>2</sub> weight % concentration <1%. On the other hand, magnetite is  
565        found attached to clay minerals of the Bodéle depression surface sediments (Moskowitz et al., 2016),  
566        which is the major source of Sahara dust in PM soil (Figure 10). Also, magnetic susceptibility

567 measurements of Sahara dust modern deposits in SE Bulgaria show a low frequency magnetic  
568 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  
569 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  
570 Mediterranean region (Rodriguez-Navarro et al., 2018). We can therefore ascribe the observed Fe-Ti  
571 oxides and (titano)magnetite an aeolian origin, from either the local igneous silicate outcrops, or the  
572 Sahara Desert. Collectively these observations imply that the magnetic enhancement of Bt compared  
573 to Bw horizon can result from a combination of fine earth illuviation of aeolian transported ultrafine  
574 magnetic particles from the Bw horizon to the Bt horizon through cryoturbation and subsequent  
575 weak pedogenetic modifications that result to the oxidation of magnetic minerals like  
576 (titano)magnetite and can also explain the reddish to yellow color hues.

577

#### 578 **5.4 Mineral weathering**

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



592 kaolinite contents can also be transported during Saharan dust transport episodes (e.g., Scheuvens et  
593 al., 2013), but specifically they are representative of the western Sahara dust provenance (PSA 1, Figure  
594 10; Rodriguez-Navarro et al., 2018). However, smectite and kaolinite are also found in modern  
595 Sahara dust samples deposited in Athens, Greece (Remoundaki et al., 2011). Therefore, we consider  
596 that the high (>80%) concentration of smectite and kaolinite in the PM soil clay (< 2  $\mu\text{m}$ ) fraction  
597 reflects the balance between direct aeolian deposition and *in-situ* weathering of detrital (aeolian  
598 and/or bedrock derived) micas and plagioclase, but the respective contributions of aeolian-  
599 transported versus that of bedrock-derived clay minerals subjected to post-depositional mineral  
600 alterations cannot be defined from the existing data.

601

## 602 **5.5 Relative timing of PM soil development**

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

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

617 and TZ cirques, the best candidates of periglacial activity that have likely resulted in the deposition of  
618 outwash sand and gravels postdate the moraine stabilization phases at  $\sim 12.5$ , 2.5, and 0.6 ka BP.

619         However, there is a 10-ka time span between the Holocene–Pleistocene boundary and the  
620 late-Holocene glacial expansions on Mount Olympus. Accepting that PM soil formation began after  
621 the moraine stabilization phase at  $\sim 12.5$  ka BP that was common to MK and TZ cirques, its  
622 production rate would be  $\sim 3 \times 10^{-5}$  m yr<sup>-1</sup> assuming that soil erosion in the low-lying PM has been  
623 minimal. This rate is considerably lower than respective soil production rates of Alpine and  
624 Mediterranean soils formed over the last 10 ka (Egli et al., 2018; Figure 8). In contrast, by considering  
625 a late-Holocene age and that the PM soil development postdates the  $\sim 2.5$  ka BP moraine  
626 stabilization phase, the soil production rate is  $\sim 1 \times 10^{-4}$  m yr<sup>-1</sup> an estimate that is in better  
627 agreement with the soil production rates presented by Egli et al. (2018) for both Alpine and  
628 Mediterranean soils. Furthermore, a late-Holocene development of PM soil broadly agrees with soil  
629 development patterns in diverse geomorphological environments in Crete (Kirsten and Heinrich,  
630 2022). If this scenario is correct, then we can further hypothesize that development of the Bt horizon  
631 could have lasted between  $\sim 2.5$  and 1.0 ka BP, before a recorded phase of intense Sahara dust  
632 accretion in Mediterranean that resulted from the combined action of an orbitally induced decrease  
633 in solar insolation and of increased aridity over North Africa (Sabatier et al., 2020). This shift could  
634 potentially explain the sharp textural boundary between the Bt and Bw horizons and the increasing  
635 Sahara dust accretion on the upper Bw horizon. The hypothesized development of the Bw horizon  
636 over the past 1 ka could have been disturbed by cryoturbation during the LIA ( $\sim 0.6$  ka BP) glacial  
637 expansion in the MK and that continues until today. Ongoing work on Mount Olympus alpine critical  
638 zone involves efforts to accurately date the MK and TZ scree interbedded layers and the PM soil  
639 profile through Optically Stimulated Luminescence dating that is aided by the high concentrations of  
640 quartz in the fine earth fraction, as well as additional geochemical analysis and estimates of the local  
641 carbonate bedrock dissolution rates and its residual geochemical composition, in an overall attempt

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

644

## 645 **6 CONCLUSIONS**

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

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

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

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

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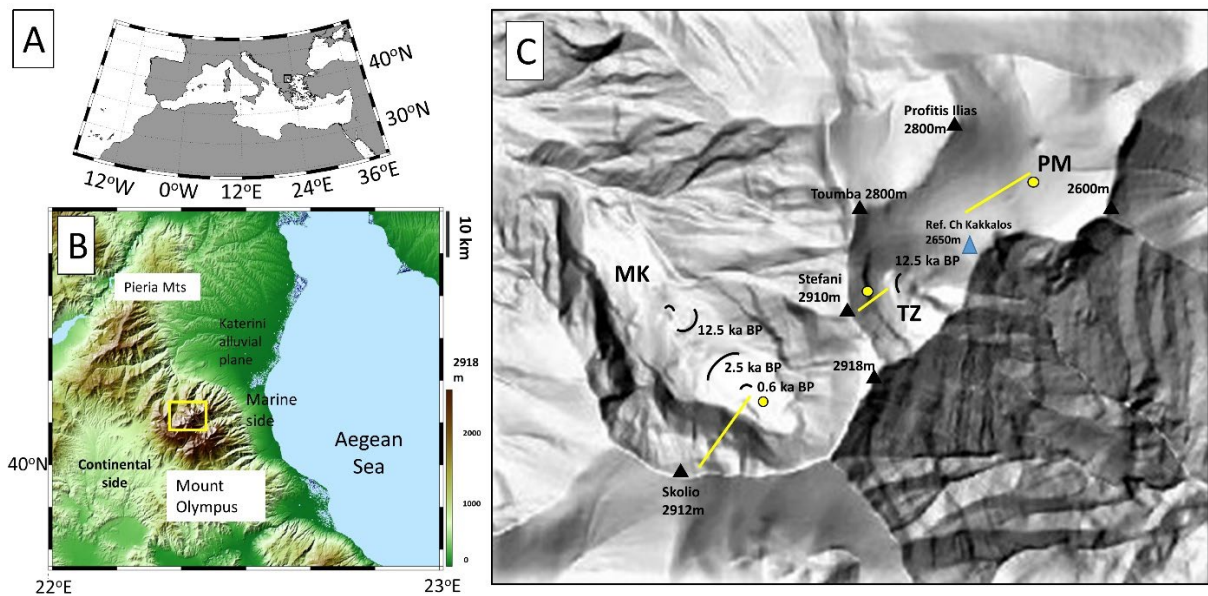
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900 **FIGURES**

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

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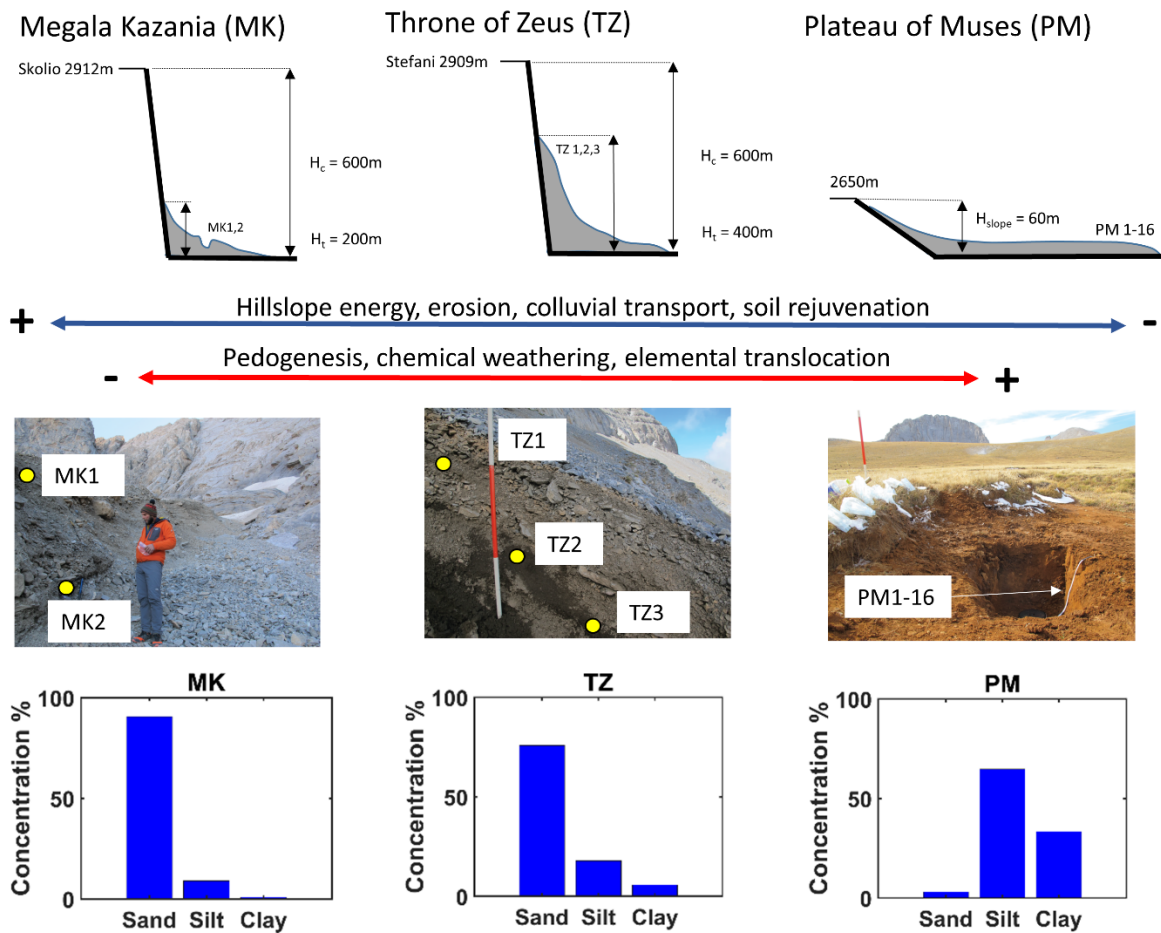


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912 **Figure 2.** Conceptual diagram of the study, with the sampling sites and their morphological profiles,  
 913 shown in Figure 1c as yellow lines, and with their respective textural characteristics that resulted  
 914 from the grain size analysis.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



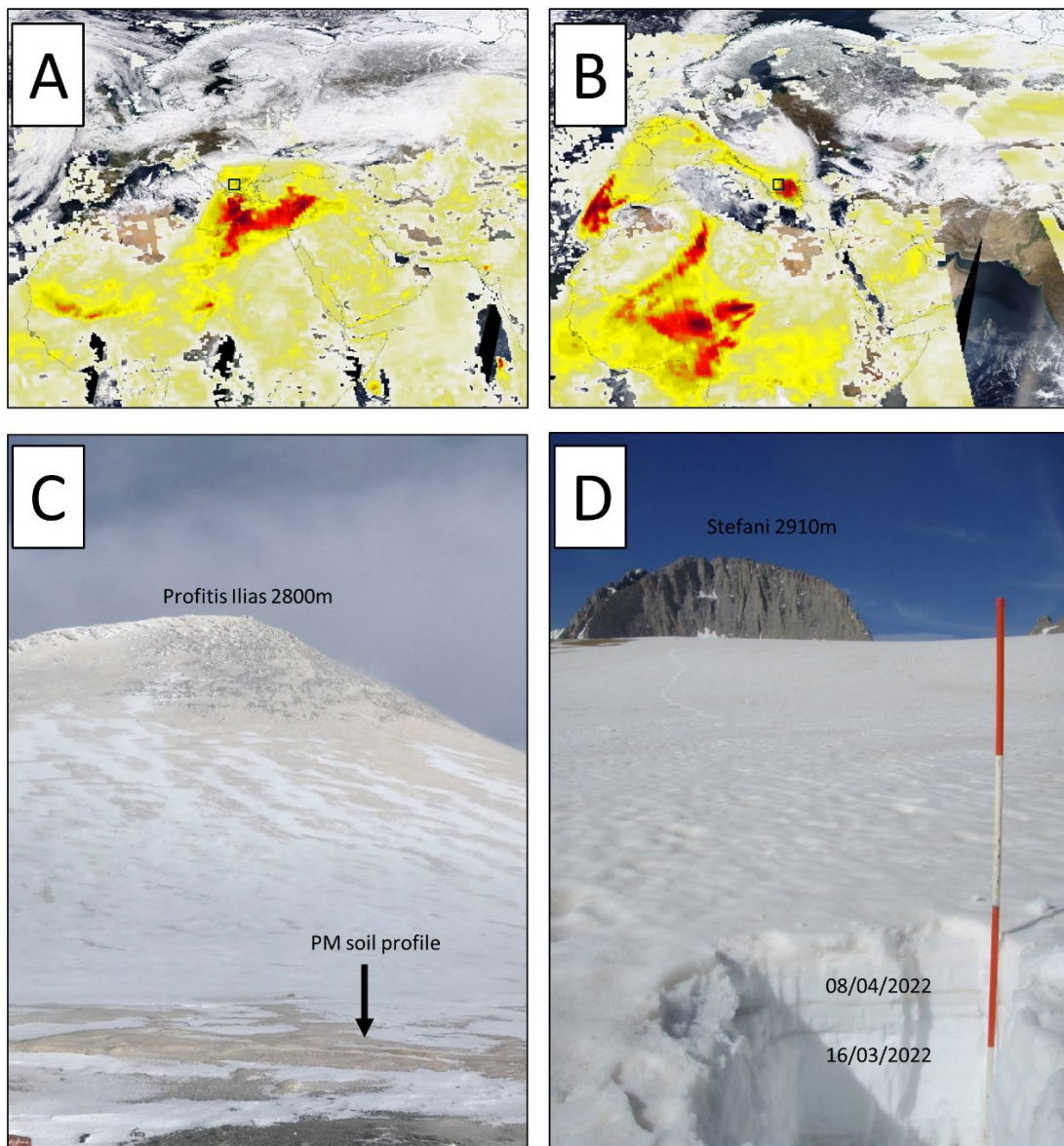
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923 **Figure 3.** Synoptic maps and direct observations of two Sahara dust episodes on Mount Olympus  
924 alpine critical zone (black rectangle). (A). Aerosol Optical Depth (AOD) during March 22, 2018, and (B)  
925 and March 16, 2022, showing the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

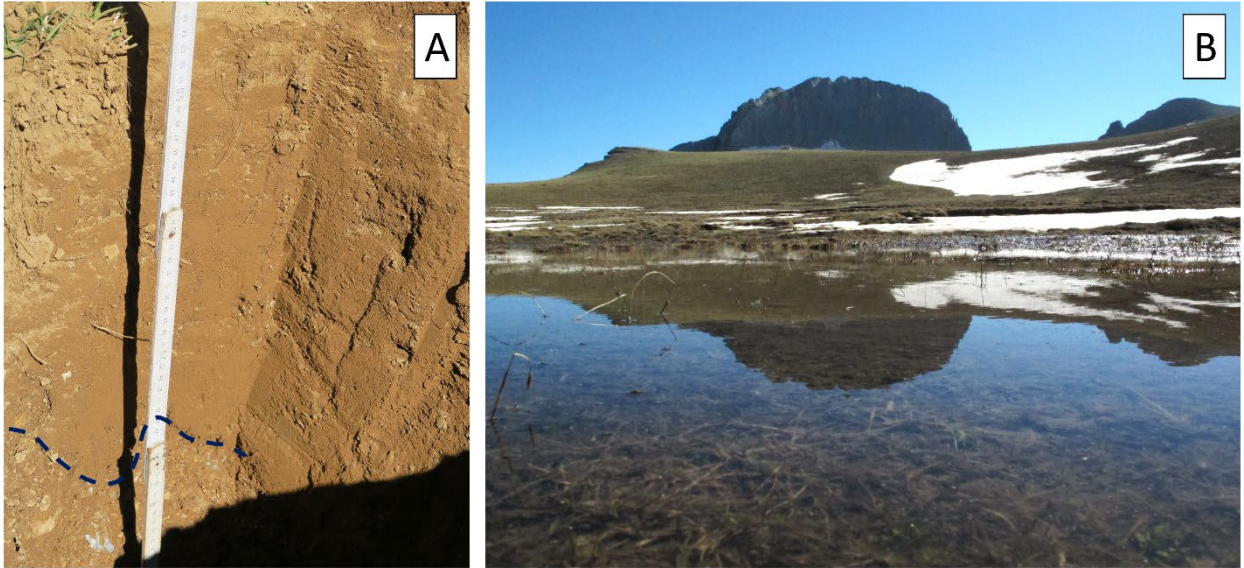
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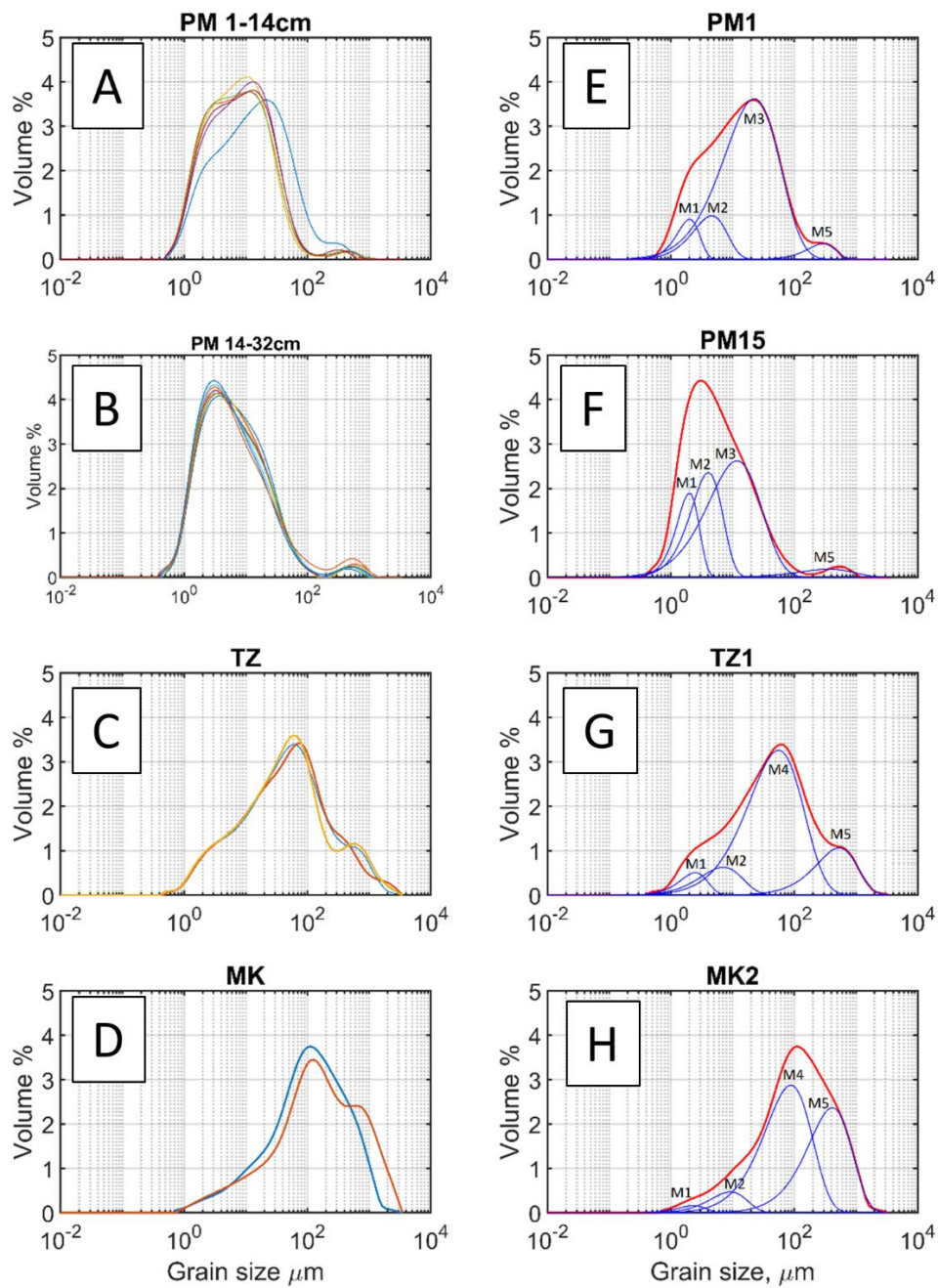
933 **Figure 4.** Evidence of soil disturbance on the Plateau of Muses under past and present-day climatic  
934 conditions. (A) An irregular gravel layer (blue dashed line) between colluvial gravel and the overlying  
935 soil resulting from cryoturbation. (B) Early summer season ongoing freeze of the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



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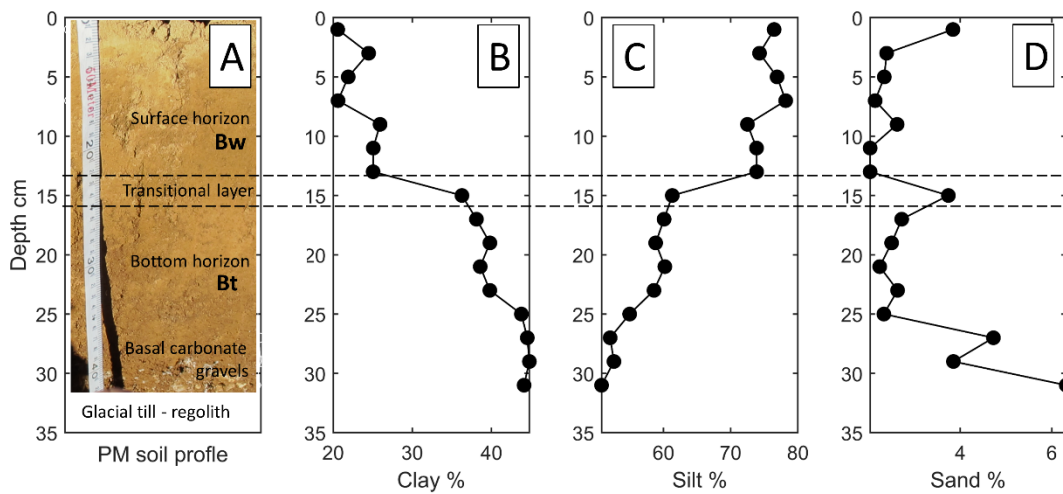
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939 **Figure 5.** Cumulative grain size distributions of the soil samples from MK, TZ and PM Bw and Bt  
 940 horizons (A, B, C and D). Surface sample PM1 (A, blue line) shows a distinct grain size distribution  
 941 from the PM soil upper layer.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



944

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

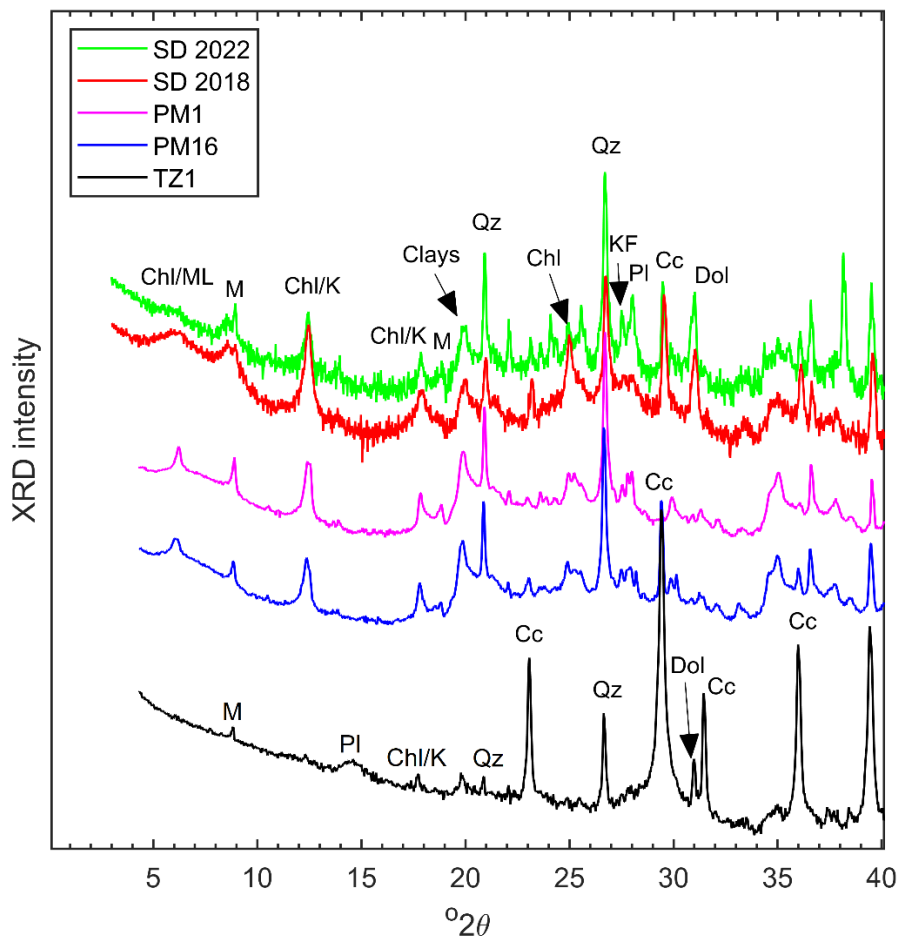


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954 **Figure 7.** X-ray diffraction patterns of soil samples PM1, PM15, TZ1 and MK2. Soils within the  
 955 hillslope high energy scree deposits are composed primarily of calcite. In contrast PM soil samples  
 956 contain quartz, clays, feldspars,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



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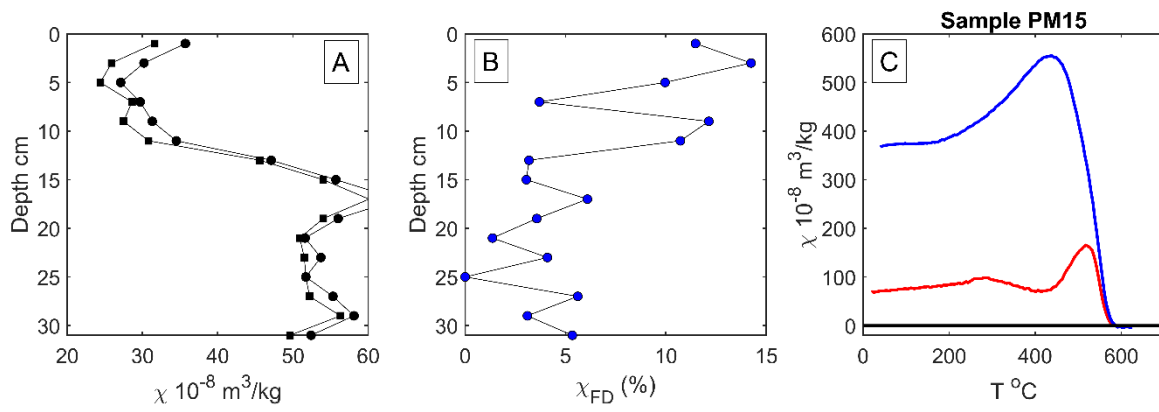
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963 **Figure 8.** Depth variations of low and high frequency (A) and frequency dependent magnetic  
964 susceptibility (B) and (C) thermomagnetic analysis results of sample PM15 (red heating curve, blue  
965 cooling curve).



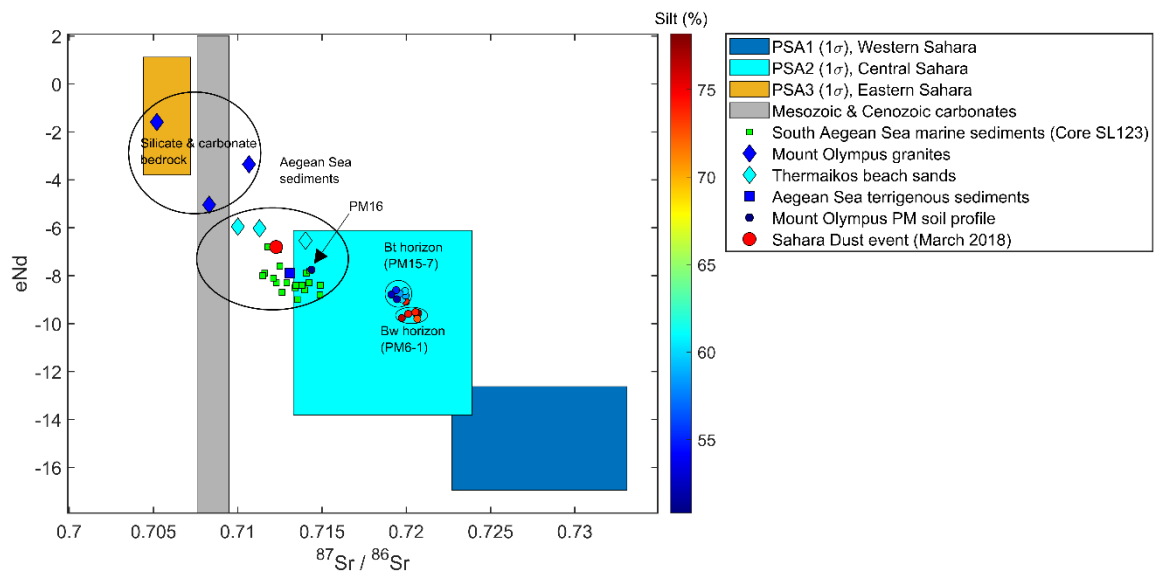
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969 **Figure 9.** Plot of  $^{87}\text{Sr}/^{86}\text{Sr}$  against  $\epsilon_{\text{Nd}}$  values of the PM soil samples with respective values of Mesozoic  
 970 and Cretaceous carbonates (Franck et al., 2021), Mount Olympus granites (located on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

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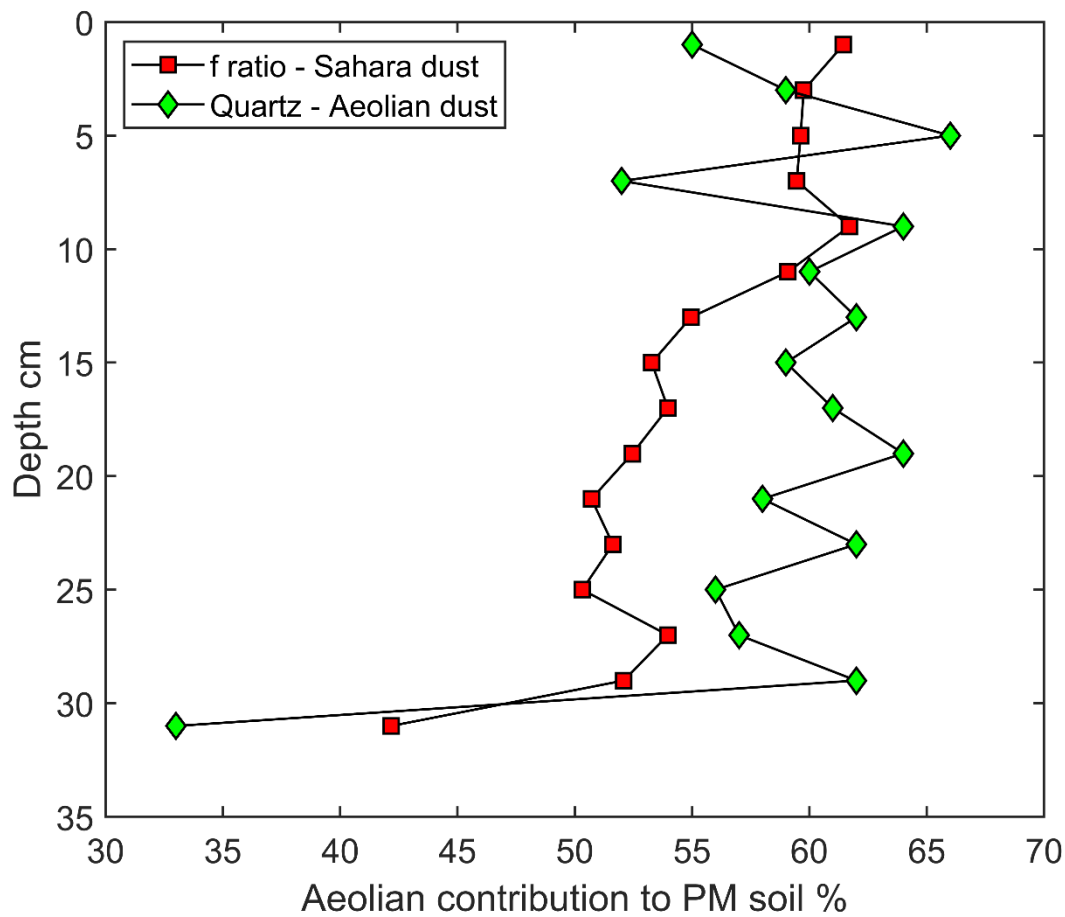
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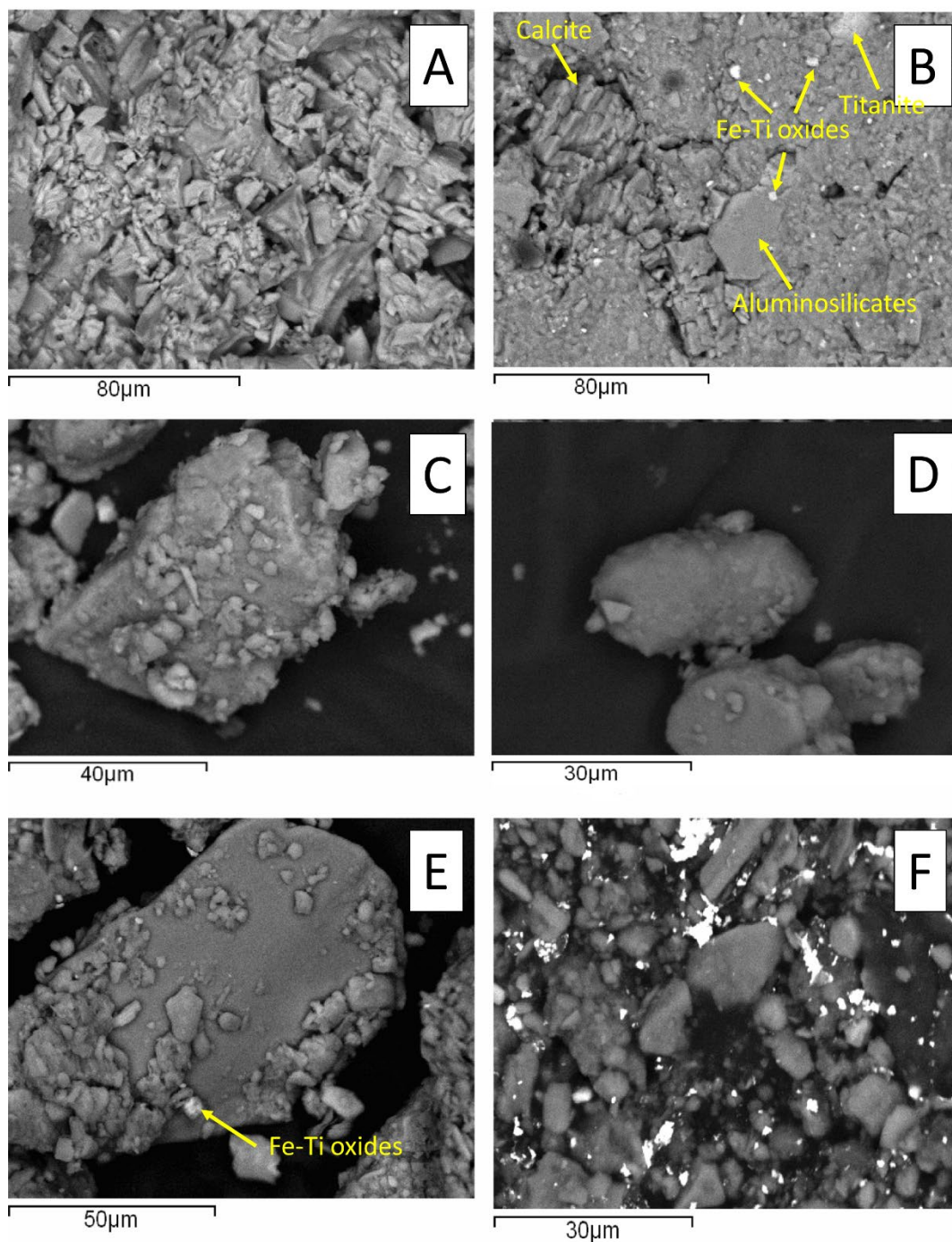
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984 **FIGURE 10.** Estimates of the relative contributions of aeolian dust accretion to PM soil as calculated  
985 by mineralogical, and isotopic proxies.



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



999

1000 **Table 1.** Physical characteristics of the soil samples retrieved from the interbedded colluvial soils of  
 1001 the Megala Kazania (MK) and Throne of Zeus (TZ) scree slopes and from the alpine soil formed on the  
 1002 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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1005 **Table 2.** Weight percent (wt %) mineralogical semi quantitative composition of the PM soil, along  
 1006 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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1009 **Table 3.** Radiogenic isotope results for the PM soil profile and the 2018 Sahara dust (SD) samples.

1010  $\epsilon$ ND values were calculated

Sample id	$^{87}\text{Sr}/^{86}\text{Sr}$		$^{143}\text{Nd}/^{144}\text{Nd}$		$\epsilon_{\text{ND}}$
	$^{87}\text{Sr}/^{86}\text{Sr}$	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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