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1 A Minoan and a Neolithic tsunami recorded in coastal sediments of 2 Ios Island, Aegean Sea, Greece

3

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9

10 Abstract

11 In this work, we document two distinct tsunami deposits on the coasts of Ios Island, Aegean Sea, Greece. The younger
12 tsunami deposit, dated 1831-1368 cal. BCE, includes both marine sediments and pumices from the ~1600 BCE Minoan
13 eruption of Santorini volcano. This is the first evidence of the Minoan tsunami in the Cycladic Islands North of
14 Santorini. Tsunami waves inundated the Manganari coastal plain, southern coast of Ios, over a distance >200 m (>2 m
15 a.s.l.). The second tsunami deposit reworks pumice from the 22 ka Cape Riva eruption mixed with marine sediment. We
16 assume a Neolithic age for this major tsunami, with a wave runup >13 m a.s.l. on the southern and eastern coasts of Ios.
17 The source of this tsunami - volcanic eruption, landslide, or earthquake - remains unknown. Additionally, we provide
18 the first on-land evidence of Cape Riva deposits outside Santorini, thus questioning previous estimates on the
19 magnitude of this eruption.

20

21 Keywords: tsunami deposits, pumice, Santorini, Minoan eruption, Cape Riva eruption, Aegean Sea

22

23

24 1. Introduction

25

26 Large explosive eruptions on volcanic islands have the potential to generate tsunami through
27 different processes such as pyroclastic flows, flank collapses, underwater explosions, volcano-
28 tectonic earthquakes, or atmospheric pressure waves (Latter, 1981; Day, 2015; Paris, 2015), as
29 demonstrated by the January 2022 eruption of Hunga Tonga - Hunga Ha'apai volcano. Volcanic
30 tsunami hazard assessment is largely based on the knowledge of past eruptions, with the most recent

31 examples being the best documented. Historical examples include tsunamis observed during the
32 eruptions of Tambora in 1815, Ruang in 1871, Krakatau in 1883, Mount Pelée in 1902, Myojin-sho
33 in 1953, Taal in 1965, Karymskoye Lake in 1996, Montserrat in 1997 and 2003, and Hunga Tonga
34 – Hunga Ha’apai in 2022 (NCEI-WDS, 2022). However, it is often difficult to infer the source of
35 these tsunamis, for the following reasons : (a) Tsunamis were generated by concurrent or combined
36 processes; (b) Volcano was little monitored (e.g. remote volcanoes, eruptions preceding the
37 instrumental period); (c) Instruments recording volcanic activity were destroyed during earlier
38 phases of the eruption; (d) Existing monitoring system was not intending for volcanic tsunami
39 detection (which is often the case). Despite the considerable progress made in the field of numerical
40 simulation, volcanic tsunamis remain difficult to model (Paris, 2015).

41 At the scale of a volcano, a tsunami generated by a large explosive eruptions is a rare event with a
42 return period of several hundreds or even thousands years. This means that the search of past events
43 is not only based on instrumental or historical records, but also on geological records. As an
44 example, Paris et al. (2014) studied tsunami deposits to reconstruct the scenario of the tsunamis
45 generated during the 1883 eruption of Krakatau. The generation of tsunami during the ~1600 BCE
46 Minoan eruption of Santorini volcano was also approached based on both geological and
47 archaeological studies (Marinatos, 1939; Doulas, 1983; Cita et al., 1984; Cita & Aloisi, 2000;
48 McCoy & Heiken, 2000; Minoura et al., 2000; Dominey-Howes, 2004; Bruins et al., 2008;
49 Goodman-Tchernov et al., 2009; Nomikou et al., 2016a; Aydar et al., 2021; Lespez et al., 2021;
50 Şahoğlu et al., 2021). Studies on tsunami deposits on the northern coast of Crete and south-western
51 coast of Turkey concluded that wave heights were exceeding 5 m and inundation runup was in the
52 order of 10 m (Minoura et al., 2000; Bruins et al., 2008; Lespez et al., 2021), depending on the local
53 topography. The tsunami is also recorded in the deep-sea sediments of the Ionian Basin as a mega-
54 turbidite (Kastens and Cita, 1981; Cita & Aloisi, 2000), and on the continental shelf off Caesarea
55 Maritima, Israel (Goodman-Tchernov et al., 2009). However, there is still a debate on the source of
56 the tsunami and its propagation outside the Aegean Sea, as illustrated by studies based on numerical
57 simulations (e.g. Pareschi et al., 2006; Novikova et al., 2011).

58 Due to the richness of its volcanic history, the Aegean Sea (Fig. 1) is a laboratory for the study of
59 volcanic tsunamis in an active tectonic setting. With 12 major explosive eruptions during the last
60 360 kyrs, Santorini volcano is the most active volcanic area of the Aegean Arc (Druitt, 1999), and
61 the iconic Minoan eruption is the only one with which a tsunami is associated. Kolumbo submarine
62 volcano, 7 km north-east of Santorini, generated a destructive tsunami in the Aegean Sea during its
63 major explosive eruption in 1650 CE (Fouqué, 1879; Nomikou et al., 2014), with tsunami deposits
64 documented on the coasts of Santorini (Ulvrova et al., 2016).

65 Earthquakes represent another source of tsunamis in the Aegean Sea (Dominey-Howes, 2002;
66 Papadopoulos et al., 2014). The region is characterised by complex extensional tectonics within the
67 African-Eurasian convergent system (Papadopoulos et al., 1986). Normal faults along the
68 extensional Anhydros and Amorgos basins (Fig. 1) have the potential to generate earthquakes with
69 a magnitude > 7 (Nomikou et al., 2016b, 2018), as demonstrated by the 1956 Amorgos earthquake
70 and tsunami (Papadopoulos & Pavlides, 1992; Okal et al., 2009).

71 However, little is known about the long-term history of tsunamis in the Aegean Sea and it is thus
72 challenging to propose credible scenarios for hazard assessment. Impact of the Minoan tsunami on
73 the Cycladic islands around Santorini is not even documented, probably because the geological
74 setting (absence of coastal marshes) and long-term human settlement do not favour the preservation
75 of tsunami deposits. In this study, we report an investigation of volcanic and tsunami deposits
76 observed on the southern and eastern coast of Ios Island, ~20 km north of Santorini (Fig. 1).

77

78 **2. Methods**

79

80 Field work and laboratory analyses focused on two sites (Fig. 1): Manganari on the southern coast
81 of Ios, and Psathi on the eastern coast. The rocky substratum is a garnet-mica schist, with marble at
82 the southern point of Manganari, and gneiss to the South of Psathi (van der Maar & Jansen, 1983).
83 The Psathi section (PS1) is located 100 m inland at altitudes ranging from 7 to 13 m. At Manganari,
84 a first section was found in the coastal plain (MA-4, 200 m inland at 2 m a.s.l.), and a second one on
85 the slopes (MA-5, 600 m inland, alt. 13 m).

86 Grain size of bulk loose samples was analysed using a particle analyser (Morphologi G3, Malvern
87 Panalytical) at LMV (Laboratoire Magmas & Volcans, France). The size and shape of pumice
88 lapilli clasts were characterised from scaled image analysis (using ImageJ software). For
89 micropaleontological analysis, samples were disaggregated using H_2O_2 and a shaking table for 24
90 hours, washed on 63 μm , 125 μm and 500 μm mesh sieves, and dried at 60°C. Approximately 100
91 benthic foraminifera specimens were counted at the stereomicroscope for each sample, but due to
92 their poor preservation these were classified mostly at the genus level. The planktonic foraminifera
93 assemblage was very limited and poorly preserved, and it was examined only qualitatively. No
94 foraminifera were observed in the 63 μm and 500 μm fractions.

95 Major element composition of pumice glass (Supplementary Table 1) was determined using an
96 Electron Probe Micro analyzer (Cameca SX100) at LMV. Whole-rock trace element composition
97 was obtained by solution Inductively-Coupled Plasma Mass-Spectrometry (ICP-MS, Agilent7500,

98 Agilent Technologies) at LMV (Supplementary Table 1). Standard-sample JA-1 was used for
99 normalizing the data.

100 X-ray tomography was realised on a 1.5 x 3 cm conglomerate sample from Manganari (EasyTom
101 RX Solutions) at LMV, with the following parameters: voltage 100 kV, current 100 μ A (power 10
102 W), and meaning of 20 images. Running a filtered back-projection algorithm on the projections
103 allows reconstructing 1435 slices with a pixel resolution of 10 μ m.

104 Dating the deposits studied here was challenging because we found very few macrofossils and
105 almost no organic content. A marine shell sampled at Manganari (Fig. 1: site MA-4) was 14 C dated
106 at the Beta Analytic laboratory, Dublin, Ireland. Calibration at 2σ was performed using the
107 Marine20 dataset (Heaton et al., 2020), assuming a DeltaR of 35 ± 70 for local reservoir correction.

108

109 **3. Results**

110

111 3.1. Stratigraphy and composition of the deposits

112

113 3.1.1. Manganari plain

114 In the Manganari plain, the rocky substratum (schist) and a brownish paleosol are covered by a 1 m-
115 thick sequence of sedimentary deposits, locally exposed on man-made trenches. Four units were
116 distinguished (Fig. 2). First unit MA-4A (135-120 cm depth) is a heterometric mixture of coarse
117 sand, subangular cobbles and pebbles from the substratum (schist) and rounded pebbles of varied
118 lithology. The deposit shows no internal organization such as clast fabric or vertical grading.
119 Second unit MA-4B (120-100 cm depth) is a medium-to-coarse sand subunit with minerals from the
120 metamorphic substratum (mostly quartz and micas), marine bioclasts (foraminifera, fragments of
121 gastropods and bivalves, rare sea-urchin spines), and rare pumice (ash size). The benthic
122 foraminifera assemblage is composed of whitish, abraded and often broken tests of the genus
123 *Ammonia*, *Asterigerinata*, *Cibicides*, *Elphidium*, *Lobatula*, and *Quinqueloculina*. Among this
124 assemblage, only few benthic foraminifera species could be determined thanks to their moderate-to-
125 good preservation grade (*Ammonia parkinsoniana*, *Lobatula lobatula*, *Elphidium crispum*). A well-
126 preserved shell of *Patella* found at a depth of 110 cm was dated $3800 \text{ BP} \pm 30 \text{ BP}$, calibrated 1862-
127 1663 BCE (1σ) or 1943-1552 BCE (2σ), with a median probability of 1754 BCE.

128 Third unit MA-4C (100-75 cm depth) is a fine-to-coarse sand (poorly-sorted) with abundant
129 minerals from the substratum), marine bioclasts, subangular-to-subrounded white pumice lapilli and
130 ash. Compared to unit MA-4-B, the marine biogenic component is finer, less abundant and poorly
131 preserved. The benthic foraminifera assemblage is composed of few tests of *Ammonia*
132 *parkinsoniana*, *Elphidium crispum*, *Lobatula lobatula*, and *Quinqueloculina* spp., often found with
133 smoothed and filled shells. Pumice lapilli are locally concentrated as well-sorted lenses in the upper
134 part of unit MA-4C (Fig. 2). Grain size of the pumice clasts ranges between 3 and 16 mm, with a
135 mean A-axis of 7 mm, thus corresponding to medium lapilli size (4-16 mm).

136 Fourth unit MA-4D (75-50 cm depth) is a fine sand with minerals from the substratum, abundant
137 pumice lapilli (fine-to-medium size) and ash (< 2 mm), with no marine bioclasts, except very few
138 sponge spicules. This facies is named “pumiceous sand” on figure 2. The sequence is overtopped by
139 50 cm of colluvium.

140

141 3.1.2. Manganari slope

142 On the north-eastern slope of Manganari, the reddish lithosol developed on the schists is locally
143 buried by patches of pumice (Fig. 3). The lowermost unit of pumice (MA-5A: 175-115 cm depth) is
144 exclusively made of subangular-to-subrounded greyish pumice lapilli, with rare mafic clasts. The
145 size of the pumice clasts ranges between 2.7 and 37 mm (i.e. from fine to coarse lapilli), with a
146 mean A-axis of 13 mm. The pumice unit is particularly well-preserved and shows a very crude
147 lamination but no vertical grading. This volcanic deposit is then eroded (see undulating contact on
148 figure 3) and reworked by a unit of alternating subunits of pumice sand and pumice gravel (unit
149 MA-5B: 115-68 cm depth) with cross-bedding in the lower part of the unit, and oblique crude
150 laminations in its upper part. Subunits are characterised by different size of pumice clasts (from ash
151 to coarse lapilli size) and different proportion of minerals from the substratum (quartz, micas,
152 garnet). In terms of texture, colour, grain size, grain shape, and mineralogy, the pumice found in
153 unit MA-5B is similar to unit MA-5A.

154 The top unit of the sequence is a pumice-rich conglomerate MA-5C (68-0 cm depth) eroding the
155 underlying deposits (unit MA-5B, see erosive contact on figure 3). The lower part of the
156 conglomerate is made of pumice lapilli (similar to unit MA-5B) and their related minerals
157 (plagioclase, pyroxene), together with minerals from the substratum, and wood. Some grains are
158 coated by a carbonated crust, as seen on SEM images (Fig. 4A). The upper part of the conglomerate
159 is characterised by the presence of same pumice as the lower part and local minerals, but there are

160 abundant marine bioclasts (foraminifera, sponge spicules, rare small molluscs, calcareous algae).
161 The benthic foraminifera assemblage is composed by whitish tiny abraded and often few broken
162 tests of the genus *Ammonia*, *Asterigerinata*, *Bolivina*, *Cibicides*, *Elphidium Haynesina*, *Lenticulina*,
163 and *Rosalina*. The conglomerate is cemented by carbonates (calcrete), forming a discontinuous
164 micritic gangue of microcrystalline calcite. X-ray tomography reveals no bedding, grain size
165 grading or grain fabric in unit MA-5C (Fig. 5).

166

167 3.1.3. Psathi

168 On the coast at Psathi, residual patches of pumice lapilli, sand and gravel are attached to the slopes
169 at altitudes ranging between 7 and 13 m (Fig. 1). The lateral variability of the sedimentary facies
170 over short distances and the calcrete cementation makes a composite stratigraphy difficult to
171 establish. However, the typical succession of facies can be summarized as follows (Fig. 6). The
172 paleosol is locally buried by a 10-30 cm-thick unit of angular-to-subangular, fine-to-coarse greyish
173 pumice lapilli (2-23 mm large, mean A-axis 8 mm). The lapilli unit is eroded by a 25-30 cm-thick
174 laminated unit of poorly-sorted very coarse sand. The sand is composed of minerals from the
175 metamorphic substratum (gneiss), together with rip-up clasts of soil (up to 6 cm large), abundant
176 marine bioclasts (Fig. 4B: benthic foraminifera, sea-urchin spines, sponge spicules), pumice lapilli
177 and pumice shards. The foraminifera are found in a moderate-to-poor state of preservation (with
178 many tests being broken and abraded), especially in the finest fraction (63-125 μm). The following
179 genus could be identified: *Ammonia*, *Asterigerinata*, *Bolivina*, *Cibicides*, *Elphidium*, *Gavellinopsis*,
180 *Haynesina*, *Lenticulina*, *Quinqueloculina*, and *Rosalina*. This first unit of sand progressively turns
181 upward to a pumice-rich facies (i.e. pumiceous sand) that still contains many marine bioclasts
182 (foraminifera, sponge spicules, sea-urchin spines, and rare ostracods). The foraminifera assemblage
183 is mostly benthic (*Ammonia*, *Asterigerinata*, *Buccella*, *Bolivina*, *Cibicides*, *Elphidium*, *Glabratella*,
184 *Gavellinopsis*, *Haynesina*, *Quinqueloculina*, and *Rosalina*), but very few badly preserved
185 planktonic foraminifera could be found (*Globigerina* and *Globorotalia* cf. *inflata*). On some
186 outcrops, it is possible to distinguish two distinct assemblages of benthic foraminifera: (1) a poorly-
187 preserved assemblage of broken and abraded tests in the finest fraction (63-125 μm); (2) and a
188 better preserved assemblage of yellowish-to-hazel colour tests in the 125 μm fraction (e.g.
189 *Asterigerinata mamilla*, *Cassidulina carinata*, *Globocassidulina subglobosa*, *Lobatula lobatula*,
190 and *Rosalina bradyi*). The upper half of the sequence displays laminated units of sand and gravel
191 that get increasingly cemented upward (Fig. 6). Their composition is similar to the underlying units,
192 but their geometry and the bedforms suggest a reworking by surface runoff.

193

194 3.2. Chemical composition of the pumices and link with explosive eruptions of Santorini volcano

195

196 The different eruptions from the Santorini-Kolumbo volcanic complex can be distinguished based
197 on their chemical composition (e.g. Druitt et al., 1999). Here we use the SiO_2 vs K_2O , TiO_2 vs K_2O ,
198 and Zr/Rb vs Ba/Y ratios to compare the composition of (1) the different pumice units found on Ios
199 Island with (2) the composition of the pumices produced by the main explosive eruptions of
200 Santorini and Kolumbo volcanoes (Fig. 7).

201 Pumices found in the lower part of the sedimentary sequences of Psathi (Fig. 6: PS-1-4) and
202 Manganari slope (Fig. 3: MA-5A) have a similar composition, and they clearly correspond to the
203 Cape Riva eruption. Indeed, they have a lower SiO_2 % wt. and K_2O % wt. but higher TiO_2 % wt.
204 compared to the Minoan eruption of Santorini and 1650 CE eruption of Kolumbo (Fig. 7). Their
205 Zr/Rb and Ba/Y ratios clearly falls in the Cape Riva field. Pumice clasts are greyish to light-brown,
206 with rare compositional zoning (banded pumice with both a rhyodacitic and an andesitic
207 component), euhedral plagioclases, pyroxenes. These characteristics are concordant with
208 descriptions of the Cape Riva products (Fabbro et al., 2014). At both sites, pumices sampled in the
209 marine sand (i.e. with marine bioclasts, e.g. PS-1-4 and PS-1-5) and marine conglomerates (e.g.
210 MA-5C) are identical to those sampled from the underlying pumice lapilli deposits (PS-1-4 and
211 MA-5A) in terms of chemical composition.

212 Pumices from the Manganari coastal plain (MA-4C) have a different composition, with higher
213 $\text{SiO}_2\%$ and $\text{K}_2\text{O}\%$ but lower $\text{TiO}_2\%$ compared to the PS-1-4 and MA-5A Cape Riva pumices. The
214 rhyodacitic composition of the MA-4C pumice (68-71% SiO_2 and ~8% $\text{K}_2\text{O}+\text{NaO}$) corresponds to
215 the Minoan eruption, except for two pumice clasts that rather belong to Cape Riva (Fig. 7). .
216 Rhyodacitic pumices have a white to light-cream colour, with very small phenocrysts of plagioclase
217 and pyroxene, as described by Heiken and McCoy (1984) on Santorini. Some pumice clasts display
218 greyish enclaves of a less-differentiated vesiculated component, which corresponds to the andesitic
219 component described by Druitt (2014) in the products of the first phase of the Minoan eruption.

220

221 4. Discussion

222

223 4.1. Implication for Santorini tephra dispersal in the Aegean

225 Inland outcrops of tephra are rare in the Cycladic Islands, except on volcanic islands themselves,
226 and the regional tephrostratigraphy is mostly based on marine tephra (Keller et al., 1978; Wulf et
227 al., 2020; Kütterolf et al., 2021). The discovery of pumice deposits on the coasts of Ios Island has
228 implication for the dispersal of tephra produced by explosive eruptions of Santorini volcano. Based
229 on their textural characteristics and chemical composition, the pumice units found on the Manganari
230 slope (Fig. 3: MA-5A unit) and at Psathi (Fig. 6: PS1-4, 140-150 cm depth) are interpreted as
231 primary pumice fall deposits formed during the Plinian phase of the Cape Riva eruption (Fabbro et
232 al., 2014), dated 22 ka (Lee et al., 2013). They are well-preserved and composed of 100% pumice
233 clasts. Crude lamination may reflect slight variations of the fallout intensity. Their grain size
234 distribution is typical of pyroclastic fall deposits (Fig. 8A).

235 The vent area of the Cape Riva eruption was located on the northern part of Santorini, between the
236 islands of Therassia and Thera, where the pumice fall deposits have a thickness of ~4 m (Druitt,
237 1985). Dispersal of the Cape Riva plume is documented through the distribution of its marine tephra
238 in deep-sea cores (Y-2 tephra in Keller et al., 1978). Thickness of the Cape Riva marine tephra
239 suggest a wide dispersal both to the North in the Aegean Sea, Sea of Marmara and Black Sea, and
240 to the East in the Mediterranean Sea (Wulf et al., 2002, 2020; Kütterolf et al., 2021 and references
241 therein) (Fig. 9A). In the near-field, Cape Riva tephra is particularly well-preserved in the Amorgos
242 and Anafi basins, with a thickness of >10 cm up to 130 km away from the volcano in the ENE
243 direction (Kütterolf et al., 2021). With a thickness of 60 cm at 24 km North of its eruptive center,
244 the Plinian deposit described here on the slopes of Manganari (Fig. 3: MA-5A unit) confirms that
245 the Cape Riva eruption strongly impacted the islands North of Santorini. This new (and first) on-
246 land estimate of the thickness of the Cape Riva pumice fall deposit outside Santorini even suggests
247 that the eruption volume and magnitude based on marine tephra thickness (~38.8 km³ and M~7.0
248 after Kütterolf et al., 2021) are slightly underestimated (Fig. 9B).

249 On the contrary, we could not find evidence of primary volcanic deposits from the Minoan eruption
250 on the coasts of Ios Island, although Minoan tephra is recorded in all marine cores around Santorini,
251 with a wide dispersal to the East (Sparks et al., 1983; Johnston et al., 2012; Kütterolf et al., 2021).
252 Pumice found in the Manganari coastal plain (Fig. 2: MA-4C) are clearly related to the first Plinian
253 phase of the Minoan eruption (Fig. 7). However, they do not correspond to a primary volcanic
254 deposit because they are mixed with a heterogeneous sand containing minerals from the local
255 substratum (quartz, micas) and marine bioclasts (foraminifera, fragments of gastropods and
256 bivalves).

257

258 4.2. A tsunami origin for the marine sands and conglomerates

259

260 Sand and conglomerate deposits found at Manganari and Psathi revealed the presence of marine
261 bioclasts (foraminifera, sponge spicules, etc.) and they display many of the diagnostic criteria
262 commonly used to identify tsunami deposits (Engel et al., 2021):

263 (1) The deposits are in discontinuity over the substratum (here, paleosol or pumice deposits) and the
264 contact is erosive. Erosion of the substratum is evidenced by the presence of rip-up clasts of soil and
265 reworked pumice.

266 (2) The deposits have a heterogeneous composition resulting from the mixing of different sources of
267 sediment (minerals from the schist and gneiss substratum mixed with pumice and marine bioclasts).
268 Benthic foraminifera assemblage is dominated by taxa related to an opened shallow water
269 environment (mostly inner-shelf, and rare outer shell species living at a water depth < 150 m). The
270 poor degree of preservation of the tests supports a reworking of foreshore or backshore deposits,
271 mixed with shoreface sediment at some locations (e.g. better preserved yellowish-to-hazel colour
272 tests at Psathi).

273 (3) Successive uprush and backwash currents produce different subunits and associated variations in
274 terms of composition (e.g. variations in the abundance of bioclasts and pumice) and grain size (from
275 gravel to sand size). As seen on, X-ray tomography images, unit MA-5C shows no bedding, no
276 grain size grading and no preserved fabric (Fig. 5). This is consistent with the en masse deposition
277 (i.e. at a very high deposition rate) of a highly-concentrated flow (Rees, 1983; Paris et al., 2020).

278 (4) The deposits are located beyond the limit of inundation of storms, which is here unequivocal
279 (PS-1 and MA-5 sites are located respectively at 7-13 m and 13 m a.s.l., and MA-4 site is located at
280 only 2 m a.s.l. but 200 m from the present-day shoreline).

281 The deposits found on the coastal plain and on the northeastern slope of Manganari (MA-4 and
282 MA-5, respectively) differ in terms of altitude and stratigraphy (Figs. 2 & 3), biogenic content
283 (more bioclasts in MA-5 compared to MA-4), and pumice composition (Fig. 7), thus suggesting that
284 they correspond to two different tsunamis, although their foraminifera assemblages are almost
285 similar.

286

287 4.3. First evidence of the Minoan tsunami in the Cycladic Islands

288

289 The 1943-1552 cal. BCE age of the *Patella* shell found in the deposits of Manganari coastal plain
290 (Fig. 2) represents a maximum age for the MA-4 tsunami deposit. This time range is compatible
291 with published ages of ~1600 BCE Minoan eruption of Santorini volcano (Friedrich et al., 2006;
292 Trevisanato, 2007; Manning et al., 2014; Şahoğlu et al., 2021), and it is thus concordant with the
293 presence of Minoan pumice in the tsunami deposit (Fig. 4A & 7). The age of the shell fits better
294 with the “high chronology” of the Minoan eruption (mid-late 17th century BCE), rather the “low
295 chronology” (16th century BCE).

296 Different successive tsunami phases are distinguished within the Minoan tsunami deposit: (1) a
297 lower sand with abundant marine bioclasts but rare pumice (Fig. 2: bioclastic sand unit MA-4B) to
298 (2) upper sands with abundant pumices and less marine bioclasts (i.e. pumiceous sand units MA-4C
299 and MA-4D). A similar succession was described by Paris et al. (2014) in the deposits of the 1883
300 Krakatau tsunami, with bioclastic sand corresponding to a pre-Plinian phase tsunami, and the
301 pumiceous sand to a Plinian phase tsunami. Pyroclastic fall deposits are interbedded between the
302 bioclastic sand and the pumiceous sand. In the case of the Minoan tsunami, the lack of preserved
303 pyroclastic fall deposits at Manganari is explained by the fact that the Plinian plume mostly
304 dispersed Eastward (Kütterolf et al., 2021 and references therein). Rare pumice ash observed in the
305 bioclastic sand unit MA-4B indicates that the eruption already started when the first tsunami
306 occurred (phase 0 of Druitt et al., 2014?), but it is difficult to infer the source of this early tsunami
307 (earthquake, landslide, atmospheric pressure waves produced by the explosions?).

308 Texture, mineralogy and chemical composition of the abundant pumices observed in units MA-4C
309 and MA-4D demonstrate that they were produced by the first Plinian phase (see section 3.2) and
310 later reworked by tsunami waves. Pumice lapilli lens observed in the upper part of unit MA-4C is a
311 kind of analogue of rounded pumice lapilli described in the deposits of the 1883 Krakatau tsunamis
312 (Carey et al., 2001; Paris et al., 2014) and 1994 Rabaul tsunami (Blong and McKee, 1995;
313 Nishimura et al., 2005). However, Minoan pumice clasts found in the Manganari tsunami deposits
314 are not completely rounded (they are subangular to subrounded, whereas the Krakatau rounded
315 pumice lapilli are subrounded to rounded), thus suggesting that they were transported over a
316 relatively short distance. A second tsunami thus impacted the coasts of Ios Island soon after the first
317 Plinian phase deposited pumice on the sea surface. Pumice accumulated along the shore as floating
318 rafts were then pushed inland by tsunami waves. Voluminous pyroclastic flows produced by the
319 third and fourth phases of the eruption are likely candidates for this second tsunami (Heiken and

320 McCoy, 1984; Druitt et al., 1999; Nomikou et al., 2016a). It is difficult to state if units MA-4C and
321 MA-4D correspond to two distinct tsunamis or two different waves during a single tsunami.

322 The scenario of the Minoan tsunami(s) at Ios Island is concordant with the one proposed by Şahoğlu
323 et al. (2021) on the western coast of Turkey, 227 km NNE of Santorini. Based on sedimentological
324 and archeological evidence, Şahoğlu et al. (2021) identified multiple tsunami events, including a
325 first (and deadly) tsunami preceding the onset of ash fall. It is difficult to infer the maximum altitude
326 (runup) reached by the waves on the coasts of Ios Island, but the large thickness (~80 cm) of the
327 tsunami(s) deposit at a present-day altitude of 2 m suggest that the tsunami inundated a large part of
328 the Manganari coastal plain, but there is no preserved deposit on the slopes around. This runup of 2
329 m represents a minimum estimate, and it cannot be concluded that tsunami waves were lower to the
330 North of Santorini (Cycladic Islands and western Turkey) compared to the South (Crete and
331 southern Turkey), where wave height probably exceeded 5 m (Minoura et al., 2000; Bruins et al.,
332 2008; Lespez et al., 2021)

333

334 4.4. A major Neolithic tsunami

335

336 Another tsunami deposit is exposed both at Manganari (MA-5) and Psathi (PS-1) at altitudes
337 ranging from 7 to 13 m, i.e. much higher than the Minoan tsunami deposit preserved in the
338 Manganari coastal plain (Fig. 1). The age of this tsunami is poorly constrained, but different
339 arguments point to a pre-Minoan Neolithic Age. First, the tsunami deposit includes pumices from
340 the 22 ka Cape Riva eruption (Fig. 7), which occurred at a time when sea level was ~110 m below
341 the present one (Satow et al., 2021). Thus, the Cape Riva eruption is not a likely source for the
342 tsunami. Second, there is a significant time lapse between the eruption and the tsunami, as
343 suggested by the accumulation of ~50 cm of slope deposits between the pumice fall deposit and the
344 tsunami deposit (Fig. 3). Third, immature calcrete observed in the MA-5C and PS-1 deposits
345 typically takes thousands of years to form in such a dry Mediterranean environment (e.g. Candy et
346 al., 2005).

347 It is difficult to infer the source of the Neolithic tsunami. Giving the absence of significant
348 explosive eruptions at Santorini volcano between the Cape Riva (22 ka) and Minoan (3.6 ka)
349 eruptions, a volcanic origin is unlikely, unless Kolumbo submarine volcano should reveal new
350 secrets. The 1650 CE Plinian eruption of Kolumbo (Fig. 1) generated a tsunami that impacted
351 Santorini, Ios, Sikinos, and even Crete (Fouqué, 1879; Nomikou et al., 2014; Ulvrova et al., 2016).

352 Surprisingly, we could not find any evidence of the 1650 CE tsunami on Ios Island, although
353 Fouqué (1879) mentioned that a local wave runup of 20 m a.s.l. associated with pumice deposition
354 was observed on a rocky shore (whose precise location is unknown). Past activity of Kolumbo
355 volcano before its 1650 CE eruption is poorly-documented, and we cannot exclude the possibility of
356 an unknown volcanic eruption and/or a mass-wasting event on its flanks. As an example, a pre-
357 Minoan unit (K3) showing chaotic and hummocky reflection on seismic data was recently
358 interpreted as a mass-wasting event (Hübscher et al., 2015).

359 A credible scenario for the Neolithic tsunami is a combined earthquake-landslide source, as
360 occurred during the 1956 Amorgos $M_s=7.4$ earthquake, which was the largest tsunamigenic
361 earthquake known in the southern Aegean Sea (Papadopoulos & Pavlides, 1992). The tsunami
362 generated wave runups up to 20 m on the southern coast of Amorgos, and 3 m on Anafi and
363 Santorini (Okal et al., 2009). Tsunami runup values and their geographic distribution are
364 incompatible with the seismic dislocation, and Okal et al. (2009) proposed that the tsunami was
365 generated by a series of landslides triggered by the earthquake.

366

367 **5. Conclusion**

368

369 The deposits of two distinct tsunamis are preserved on the coasts of Ios Island. The youngest
370 deposit corresponds to the Minoan tsunami generated by the eruption of Santorini volcano ~3.6 ka
371 ago. We thus confirm that the tsunami also propagated North of Santorini Island, and not only to
372 Crete or Turkey. Tsunami runup was higher than 2 m on the southern coast of Ios Island. The
373 second deposit, proposed as Neolithic aged, records a newly discovered tsunami event in the
374 Aegean, with a magnitude larger than the Minoan tsunami. The source of the Neolithic tsunami is
375 unknown.

376 This work confirms the relevance of paleotsunami studies for extending the tsunami record and
377 improving hazard assessment. Without the support of geology through the identification of a major
378 Neolithic tsunami, the Minoan and the 1956 tsunamis could be considered as worst-case scenarios,
379 which is obviously not the case. Following the 2011 Tohoku-oki tsunami in Japan, The Japanese
380 government voted a new law for tsunami disaster prevention plans in December 2011, and
381 recommended that local governments prepare for the maximum possible tsunami, based on analysis
382 of ancient documents and tsunami deposits (Goto et al., 2014).

383 This study also underscores the complexity of tsunami generation during major explosive eruptions,
384 such as the Minoan eruption of Santorini, the 1650 eruption of Kolumbo, the 1883 eruption of
385 Krakatau, and the 2022 eruption of Hunga Tonga – Hunga Ha’apai. Although these four eruptions
386 each have a particular succession of events, they all demonstrate the diversity of volcanic tsunamis
387 in terms of:

- 388 • Timing: Tsunamis may happen not only during the Plinian paroxysm, but also during the
389 early sub-Plinian phase or the waning phase.
- 390 • Source mechanisms: These volcanic tsunamis can be generated by a range of phenomena
391 such as earthquakes, landslides, underwater explosions, pyroclastic flows, eruptive column
392 collapse, or atmospheric pressure waves.
- 393 • Space: Depending on their source and the bathymetry, volcanic tsunamis might have a local,
394 regional, or a global impact, as demonstrated by the recent events in Tonga (Lynett et al.,
395 2022; Omira et al., 2022).

396

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405

406 **Data Availability**

407 All data generated or analysed during this study are included in this article (and its supplementary
408 information files) or available from the corresponding author on reasonable request.

409

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589 **Figure captions**

590
591 Fig. 1 - Location map of study sites on Ios Island, Aegean Sea, Greece.

592
593 Fig. 2 - Sedimentary log of the Manganari plain (MA-4 site) and microscope views. Facies s:
594 sand; pl: pumice lapilli; ps: pumiceous sand; co: colluvium.

595
596 Fig. 3 - Sedimentary log of the Manganari northeastern slope (MA-5 site) and field views.
597 Facies pl: pumice lapilli; ps: pumiceous sand; pg: pumice gravel; pc: pumice conglomerate.

598
599 Fig. 4 - SEM images of (A) a pumice-rich tsunami deposit at Manganari (MA-5C unit on fig. 3),
600 with pumice clasts in yellow, and marine bioclasts in blue; and (B) a bioclast-rich tsunami deposit
601 at Psathi (sampled at 90 cm depth on PS1-5 section, Fig. 6).

602
603 Fig. 5 – X-ray computed tomography images (CT-scan) of MA-5C unit. A: original slices in
604 grayscale (note heavy minerals in white); B: color map of a single slice, with heavy minerals in red
605 (e.g. garnet), other high-density minerals in yellow (e.g. carbonates), silicates in light blue (e.g.
606 quartz), and low-density pumice clasts in dark blue. C: reconstructed volume of the sample; D:
607 examples of colored slices through the sample.

608
609 Fig. 6 – Sedimentary logs of the Psathi site (PS). Facies pl: pumice lapilli; lpl: laminated pumice
610 lapilli; ps: pumiceous sand; pg: pumiceous gravel; pc: pumice conglomerate; ; s: sand; lsg:
611 laminated silt-to-gravel; cal: calcrete; co: colluvium.

612
613 Fig. 7 - Chemical composition of pumice fall deposits and pumice in tsunami deposits on Ios
614 Island, compared to the composition of pumice produced by the main explosive eruptions of
615 the Santorini-Kolumbo volcanic complex, including data from Druitt et al. (1999), Simmons
616 et al. (2016) for Lower Pumice 1, Gertisser et al., (2009) and Keller et al. (2014) for Lower

617 Pumice 2, Fabbro et al. (2013) for Cape Riva, Zellmer (1998) for the Late Bronze Age
618 eruption and Cantner et al. (2014) for Kolumbo.

619

620 Fig. 8 – Grain-size and morphological characteristics of pumice lapilli clasts. A: median size
621 Md (ϕ) vs. sorting σ (ϕ) (pyroclastic flow and pyroclastic fall fields after Walker, 1971); B:
622 circularity vs. solidity.

623

624 Fig. 9 – Isopach maps of the Cape Riva (22 ka) tephra. Italic numbers in blue indicate
625 thickness (cm). A: regional isopach map (Kütterolf et al., 2021); B: isopach map from
626 Santorini (Druitt, 1985) to Ios Island (this work).

627