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Raphaël Paris[1], Alessandra Smedile[2], Simon Falvard[1], Jean-Luc Devidal[1], Krzysztof Suchorski[1]

[1] Université Clermont Auvergne, CNRS, IRD, OPGC, Laboratoire Magmas et Volcans, F-63000 Clermont-Ferrand, France

[2] Istituto Nazionale di Geofisica e Vulcanologia, Rome, Italy.

Abstract

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

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

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1. Introduction

Large explosive eruptions on volcanic islands have the potential to generate tsunami through different processes such as pyroclastic flows, flank collapses, underwater explosions, volcano-tectonic earthquakes, or atmospheric pressure waves (Latter, 1981; Day, 2015; Paris, 2015), as demonstrated by the January 2022 eruption of Hunga Tonga - Hunga Ha’apai volcano. Volcanic tsunami hazard assessment is largely based on the knowledge of past eruptions, with the most recent examples being the best documented. Historical examples include tsunamis observed during the eruptions of Tambora in 1815, Ruang in 1871, Krakatau in 1883, Mount Pelée in 1902, Myojin-sho in 1953, Taal in 1965, Karymskoye Lake in 1996, Montserrat in 1997 and 2003, and Hunga Tonga – Hunga Ha'apai in 2022 (NCEI-WDS, 2022). However, it is often difficult to infer the source of these tsunamis, for the following reasons : (a) Tsunamis were generated by concurrent or combined processes; (b) Volcano was little monitored (e.g. remote volcanoes, eruptions preceding the instrumental period); (c) Instruments recording volcanic activity were destroyed during earlier phases of the eruption; (d) Existing monitoring system was not intending for volcanic tsunami detection (which is often the case). Despite the considerable progress made in the field of numerical simulation, volcanic tsunamis remain difficult to model (Paris, 2015).

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

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

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

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

2. Methods

Field work and laboratory analyses focused on two sites (Fig. 1): Manganari on the southern coast of Ios, and Psathi on the eastern coast. The rocky substratum is a garnet-mica schist, with marble at the southern point of Manganari, and gneiss to the South of Psathi (van der Maar & Jansen, 1983). The Psathi section (PS1) is located 100 m inland at altitudes ranging from 7 to 13 m. At Manganari, 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 the slopes (MA-5, 600 m inland, alt. 13 m).

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

Major element composition of pumice glass (Supplementary Table 1) was determined using an Electron Probe Micro analyzer (Cameca SX100) at LMV. Whole-rock trace element composition was obtained by solution Inductively-Coupled Plasma Mass-Spectrometry (ICP-MS, Agilent7500, Agilent Technologies) at LMV (Supplementary Table 1). Standard-sample JA-1 was used for normalizing the data.

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

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

3. Results

3.1. Stratigraphy and composition of the deposits

3.1.1. Manganari plain

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

Third unit MA-4C (100-75 cm depth) is a fine-to-coarse sand (poorly-sorted) with abundant minerals from thesubstratum), marine bioclasts, subangular-to-subrounded white pumice lapilli and ash. Compared to unit MA4-B, the marine biogenic component is finer, less abundant and poorly preserved. The benthic foraminifera assemblage is composed of few tests of *Ammonia parkinsoniana*, *Elphidium crispum*, *Lobatula lobatula*, and *Quinqueloculina* spp., often found with smoothed and filled shells. Pumice lapilli are locally concentrated as well-sorted lenses in the upper part of unit MA-4C (Fig. 2). Grain size of the pumice clasts ranges between 3 and 16 mm, with a mean A-axis of 7 mm, thus corresponding to medium lapilli size (4-16 mm).

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

3.1.2. Manganari slope

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

The top unit of the sequence is a pumice-rich conglomerate MA-5C (68-0 cm depth) eroding the underlying deposits (unit MA-5B, see erosive contact on figure 3). The lower part of the conglomerate is made of pumice lapilli (similar to unit MA-5B) and their related minerals (plagioclase, pyroxene), together with minerals from the substratum, and wood. Some grains are coated by a carbonated crust, as seen on SEM images (Fig. 4A). The upper part of the conglomerate is characterised by the presence of same pumice as the lower part and local minerals, but there are abundant marine bioclasts (foraminifera, sponge spicules, rare small molluscs, calcareous algae). The benthic foraminifera assemblage is composed by whitish tiny abraded and often few broken tests of the genus *Ammonia, Asterigerinata, Bolivina, Cibicides, Elphidium Haynesina, Lenticulina,* and *Rosalina.* The conglomerate is cemented by carbonates (calcrete), forming a discontinuous micritic gangue of microcrystalline calcite. X-ray tomography reveals no bedding, grain size grading or grain fabric in unit MA-5C (Fig. 5).

3.1.3. Psathi

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

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

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

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

Pumices from the Manganari coastal plain (MA-4C) have a different composition, with higher SiO_2% and K_2O% but lower TiO_2% compared to the PS-1-4 and MA-5A Cape Riva pumices. The rhyodacitic composition of the MA-4C pumice (68-71% SiO_2 and ~8% K_2O+NaO) corresponds to the Minoan eruption, except for two pumice clasts that rather belong to Cape Riva (Fig. 7). . Rhyodacitic pumices have a white to light-cream colour, with very small phenocrysts of plagioclase and pyroxene, as described by Heiken and McCoy (1984) on Santorini. Some pumice clasts display greyish enclaves of a less-differentiated vesiculated component, which corresponds to the andesitic component described by Druitt (2014) in the products of the first phase of the Minoan eruption.

4. Discussion

4.1. Implication for Santorini tephra dispersal in the Aegean

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

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

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

4.2. A tsunami origin for the marine sands and conglomerates

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

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

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

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

(4) The deposits are located beyond the limit of inundation of storms, which is here unequivocal (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 only 2 m a.s.l. but 200 m from the present-day shoreline).

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

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

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

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

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

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

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

4.4. A major Neolithic tsunami

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

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

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

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

5. Conclusion

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

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

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

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

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Data Availability

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

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Figure captions

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

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

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

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

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

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

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

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

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

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