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Large seafloor rupture caused by the 1956 Amorgos tsunamigenic earthquake, Greece

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In the Mediterranean Sea, the probability that a large earthquake-triggered tsunami will occur in the coming decades is high. Historical tsunami database informs us on their geographical occurrence but their sources, i.e., the faults that slipped during earthquakes and displaced the seafloor to generate tsunamis, are often unknown. Here we identify the submarine rupture of the Amorgos earthquake that on July 9, 1956, triggered the largest mediterranean tsunami in the past two centuries. Using submarines, we explored major normal faults in the epicentral area, and discovered a large surface rupture along the 75-km long Amorgos fault. The 9.8–16.8-m large seafloor offset is compatible with a Mw7.5 event. This finding prompts a reassessment of the largest (≥ 20 m) tsunami wave origin, previously attributed to earthquake-triggered submarine mass-wasting. It demonstrates that tsunami source can be determined several decades after an event, a key information to better assess future seismic and tsunami hazards.

Surface ruptures generated by large continental earthquakes, whether historical^{1,2} or recent³, are today systematically mapped through fieldwork, satellite data, and high-resolution topographic analyses. They provide key information that allows the seismic hazard of a region to be evaluated, in particular through the identification of the causative fault, the extent of the rupture, and the amount of slip generated by the earthquake. Such work was performed in only a few weeks following the 2023 Turkey–Syria earthquake⁴ for example. When the earthquake occurs offshore, similar analyses are challenging to perform, but they are of prime importance to understand the triggering of tsunamis and evaluate the seismic hazard along submarine faults⁵. The recent deployment of submarine vehicles (such as Remotely Operated Vehicles ROVs), however, offers new opportunities to image undersea fault scarps^{6–9}, identify fault planes that have been recently exhumed by slip during an earthquake¹⁰, and map and quantify the related surface rupture¹¹. With such a vehicle, we investigated the faults around the proposed epicenter of the 1956 Amorgos earthquake to identify seafloor ruptures and determine the fault responsible for this event.

The 1956 earthquake and tsunami

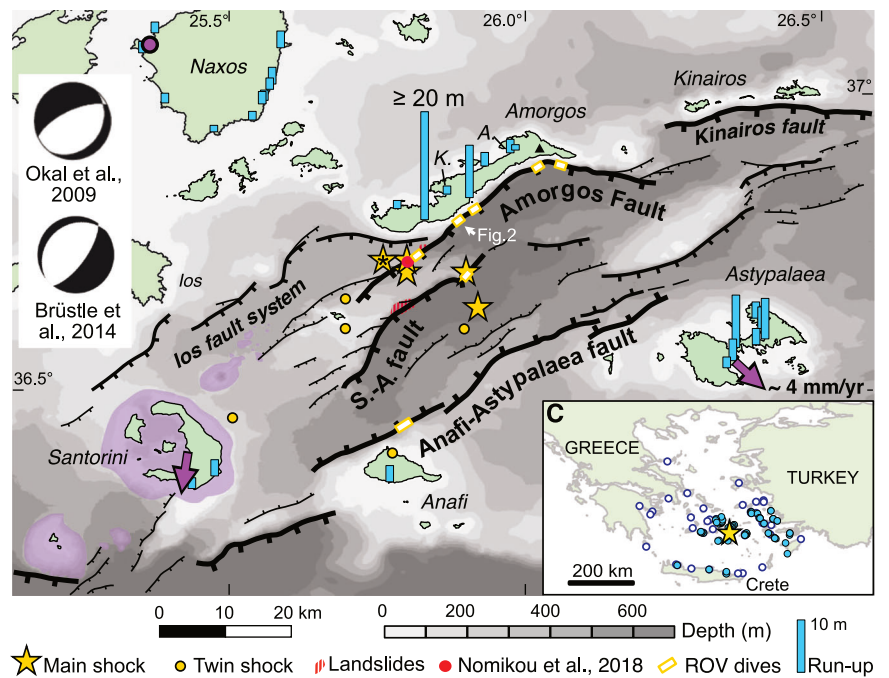
The Amorgos earthquake occurred on July 9, 1956, offshore Santorini and Amorgos (Cyclades, South Aegean Sea), along the Hellenic Volcanic Arc. It

was recorded by a small number of seismometers, enabling seismologists to determine a magnitude of 7.2–7.8¹², depending on the authors. Several epicenters were obtained¹³ that located the earthquake between 5 and 20 km south of Amorgos island (Fig. 1). Its hypocentral depth, recognized to be poorly constrained¹⁴ and debated¹⁵, varies between 10 km and 45 km and was recently re-evaluated at ~ 25 km¹³, that is, at the Moho depth of the Hellenic arc¹⁶. The main shock was followed by a series of aftershocks. The first aftershock, called the twin earthquake, had a magnitude estimated between 6.0 and 7.2, and occurred only 13 min later, closer to Santorini (Fig. 1) and probably deeper (40–95 km), possibly along the subduction plate boundary¹³. The main shock caused severe damage to the surrounding islands, subsequently enhanced by the twin shock, especially in Santorini¹⁷. More than 3200 buildings were damaged, including ~ 500 that were completely destroyed; in addition, 54 people were killed in Santorini, and 100 people were injured.

Several focal mechanisms were obtained to characterize the faults that are likely to have broken. The best-constrained ones point to a NE–SW striking normal fault plane^{13,14}, in agreement with the local tectonic context and secondary faults visible along the southern coast of the island^{17–19} (Fig. 1). The coseismic slip probably occurred on a SE-dipping plane, calculated to be either a low-dipping plane (25°)¹⁴ or a steeper dipping plane

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Fig. 1 | Seismotectonic map in the epicentral area of the July 9th, 1956 earthquake. Several epicenters for the main shock (star) and twin shock (dot) are represented, as summarized in Brüstle et al.¹³, with two proposed focal mechanism^{13,14} calculated for the epicenter marked with a star and an asterisk. Main faults are represented by black lines, with thicker traces for longer and taller faults, and are modified from previous works^{19,29,33}. S.-A. fault: Santorini-Amorgos fault. Portions of faults explored with the ROVs during the AMORGOS-23 cruise are localized by yellow and white rectangles. Submarine landslides identified²⁵ are in dashed red, while the observation¹⁹ of a probable fault mirror associated with the 1956 event along the Amorgos fault is a red dot. Measured run-ups¹⁴ of the 1956 tsunami are represented as blue bars. In the inset, other tsunami observations of the 1956 event are located by blue¹⁴ and white²⁰ dots. Light purple areas are onshore and offshore volcanoes. The purple arrows on Astypalaea and Santorini show their relative displacements southeastward and southwestward, respectively, at a rate of ~4 mm/yr, with respect to central Aegean (Naxos)³¹. Mount Kroukelos on Amorgos is denoted by a black triangle. The two archeological sites that demonstrate long-term subsidence of the northern coast of Amorgos¹⁸ are located in Katapola (K.) and Aegiali (A.).



(~65°)¹³ (both shown in Fig. 1), the latter being more typical of a normal fault. However, our knowledge of the fault that broke and produced this earthquake has until now been incomplete, due to our inability to identify and map the undersea fault.

In addition, the main shock triggered a large tsunami that reached the coastlines of Crete, Peloponnesus, western Turkey^{14,20,21}, and that was recorded by one tide gauge in Israel²². Amorgians today still recall that after the earthquake, the sea retreated, allowing pedestrians to cross the Katapola Bay, before it inundated the land, with waves of 2–3 m high (Fig. 1). Along the southern coast of Amorgos island, run-ups of up to 20 m were reported^{14,20} while the northern coastline of Astypalaea island was also flooded with a run-up of up to 10 m^{14,20,23} (Fig. 1). These values are the highest reported in the Mediterranean basin in the twentieth and twenty-first centuries²⁴. Despite a re-assessment of the testimonies of this tsunami¹⁴, the details of the arrival time of the wave remain vague¹⁴ across the archipelago. The polarity of the first wave is not well established²¹, while tide gauges installed in Crete and in Leros were damaged by the earthquake, so the tsunami arrival was not recorded²¹ in the near-field.

To explain the variation and local tsunami run-up heights along the coasts, a second source of sudden submarine seafloor motion was proposed, in addition, to slip on a low-angle normal fault¹⁴. Waves were probably enhanced by submarine landsliding²⁰, later observed at the seafloor²⁵ in the archipelago (Fig. 1). Such a scenario was tested with tsunami modeling^{14,22} so as to reproduce tsunami data (run-ups only¹⁴), using pre-determined source geometries. The geometry of the basins and the coastal bathymetry that greatly influence the run-up values at the coasts²⁶ were, and are mostly still, unknown and thus unused in tsunami modeling¹⁴. All in all, the primary source of the earthquake and tsunami remains debated²⁷, as epicentral solutions, tsunami data, and coastal bathymetry data are not sufficiently constrained.

On-land investigations were conducted to gain new insights into the origin of the events. Historical aerial images of Amorgos shorelines, acquired before and after the 1956 earthquake, suggest a coseismic uplift of the southern coast, while archeological masonries nowadays lying below sea level indicate long-term subsidence of the island's northern coast¹⁸ (Fig. 1). These vertical motions are compatible with the break of a steep normal fault located within 5 km of the Amorgos' southern coast¹⁸. Conversely, far-field

subsidence of Holocene shorelines²⁸ are compatible with the co- and post-seismic motion generated by a deep and low-angle normal fault. Using the different markers visible on land and in coastal areas is clearly insufficient to understand this earthquake and tsunami, as they do not converge towards a common source.

The submarine faults offshore Amorgos

In the past 10 years, marine geophysical data were acquired^{19,25,29,30} with sufficient resolution to identify, map, and characterize the faults in the area (Fig. 1). A 750-m deep, NE-SW striking trough, measuring 35 km in width, exists between the islands of Ios, Amorgos and Kinairos to the north, and Anafi and Astypalaea to the south. This trough is bordered to the north by a set of NE-SW to E-W striking normal faults, dipping to the south-southeast.

The main fault is the Amorgos fault, which constitutes the southern cliff of Amorgos island, whose summit reaches 821 m (Mt Kroukelos, Fig. 1). The main segment of the fault measures 45 km (Fig. 1, thick dark line), and up to 75 km if we consider secondary structures that connect or align with the main fault at its tips (Fig. 1, thin dark lines). Its activity induces the subsidence of its hanging wall, marked by the depth of Amorgos basin, reaching ~750 m below sea level^{19,29}. Seismic reflection profiles acquired in the basin reveal that 700 m of sediments, tilted to the north, i.e., toward the fault, cover the alpine basement¹⁹, indicating about 2.2 km of vertical offset by the Amorgos fault.

The southern border of the trough is structured by the NE-SW striking, NW-dipping Anafi-Astypalaea fault system that extends from NW of Anafi to NW of Astypalaea island (Fig. 1). This fault is segmented, and in its central part it presents several sub-parallel smaller faults offsetting its hanging wall²⁹. Its cumulative offset is lower than that of the Amorgos fault, about 1 km in its western part³⁰, and it measures about 65 km in length.

In between the two main antithetic faults, several other faults offset the seafloor, creating horsts and grabens, including the Santorini-Amorgos (S-A) fault, located at ~8 km from the southern coasts of Amorgos. This structure is an NE-SW striking, SE-dipping normal fault, offsetting the basement by about 1.1 km¹⁹. In total, the Santorini-Amorgos fault measures about 55 km.

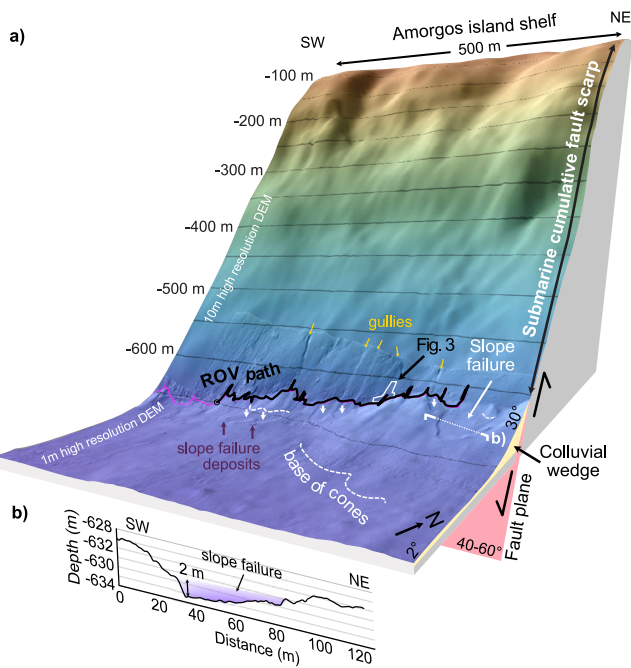


Fig. 2 | 3D view of the morphology of the submarine Amorgos fault. **a** A 500 m long portion of the fault (located in Fig. 1) is represented in 3D with no vertical exaggeration. The HROV investigated the base of the fault scarp (navigation in pink and black). Gullies (yellow arrows) incise the fault mirror. At their outlets, cones of sediments (dashed white lines) are visible, they are affected by slope failures (white arrows) that deposit in the deep basin. **b** Bathymetric profile across one slope failure allows determining that it is at most 2 m deep (purple area).

The orientation of this fault system is compatible with the two nodal planes of the focal mechanisms of the 1956 earthquake (Fig. 1). At the scale of the trough, these faults thus seem to accommodate mainly an NW–SE extension, in line with recent GPS data showing an NW–SE-oriented velocity gradient of 4 mm/yr between Naxos and Astypalaea³¹ (Fig. 1). On a larger scale, this fault system might accommodate some lateral motion of the southern Aegean domain with respect to the central Aegean domain, but this is still debated^{29,31–33}.

Considering the uncertainty of the earthquake location and the incompleteness of the tsunami data, the three main faults of this system, that is, the Amorgos, Santorini–Amorgos, and Anafi faults, are all good candidates as sources for the 1956 main shock (Fig. 1). Moreover, as the surface (and thus length) of a fault is proportional to the magnitude of the largest earthquake it can generate^{34,35}, the three faults are all long enough to host an $M \geq 7$ earthquake such as the Amorgos 1956 main shock. However, recently acquired seismic reflection profiles¹⁹ imaged a particularly steep 8–10 m high scarp at the base of the Amorgos fault (Fig. 1). This suggests that a recently exhumed fault mirror may be preserved here, and thus that this fault could be the source of the 1956 earthquake.

We surveyed the three faults for the first time using an autonomous underwater vehicle (AUV) and a hybrid remotely operated vehicle (HROV) onboard the R/V Europe^{36,37}, in order to characterize the faults' morphologies with bathymetric data (resolutions of 1 and 10 m), and image its surface to identify potential ruptures using 4 K video imagery (Fig. 1). The strategy carried out to survey optically each of the three faults was the same: HROV dived at fault sites showing the steepest cumulative scarp (identified in the bathymetry), and the simplest geometry (deformation is accommodated along one fault scarp only, observations were made far from fault relays). Dives were performed close to the epicentral area, away from large mass-wasting scars and associated deposits that could degrade or cover the markers of surface rupture on the seafloor. With this strategy, we identified that the Amorgos fault is the only one to exhibit a freshly exhumed fault mirror at the base of its cumulated scarp (Figs. 2 and 3).

Results: a fault mirror with fresh exhumation traces

Offshore of the south-eastern coast of Amorgos island, the new high-resolution bathymetric data reveals that the Amorgos fault exhibits a >600 m high submarine scarp visible all along its strike. At the HROV dive site presented below, this scarp is particularly steep (50–60°) and linear (Fig. 2a, Supplementary Fig. S1). It shows neither segmentation nor synthetic or antithetic parallel splays, indicating that the deformation localizes here along a single fault plane.

In the 1-m resolution digital elevation model (DEM), the base of the cumulative scarp is smooth, showing only a few narrow gullies parallel to the slope that incises it (Fig. 2a). At the gullies' outlet, small cones are visible, and cover the fault scarp locally. The base of the fault scarp dips between 40° and 60° (Fig. 2a, Supplementary Fig. S2), a range comparable to the dips determined at depth with geophysical¹⁹ and seismological²⁷ observations (38°–66°). A colluvial wedge covers its foot and dips toward the basin, with slopes up to 30° close to the scarp, and only 2° at 300 m far from the scarp (Figs. 2a and S2). Close to the scarp, the colluvial wedge has been locally eroded by failures and gravity collapses and shows local depressions that are up to three meters deep (Figs. 2a, b and S1).

Using the 4 K camera of the HROV Ariane, we imaged both the fault scarp and the colluvial wedge (Fig. 3, Supplementary Movie 1). The surface of the colluvial wedge is covered by unconsolidated fine sediments that are easily mobilized by fishes, currents, and the HROV propellers. Below the surface, the colluvial wedge exhibits stratified sediments exposed at the edges of the depressions that dissect the wedge (Fig. S1). As the edges of these failures are vertical, sediments forming the first few meters of the wedge are thus cohesive and probably indurated. Strata are ~1 m thick, composed of fine light-colored sediments, and dipping toward the basin (away from the fault).

The fault scarp is made of indurated rocks whose nature and geometry could not be determined during the dives. We speculate that they are similar to the metamorphic rocks outcropping on land, and forming the Alpine basement offset by the fault¹⁹. We identified that the base of the cumulative fault scarp displays a smooth and well-preserved surface (Fig. 3, Supplementary Fig. S3, Supplementary Movie 1), that is continuous over at least a length of ~500 m. This surface exhibits numerous striae that are in the dip direction of the fault plane. Some of them are continuous over the colluvial wedge indicating that they are rills linked to sedimentary and erosion processes (Figs. 3a and S3f). Other striae are covered at their lower end by the colluvial wedge (Fig. S3b and Supplementary Movie 1), suggesting that they are not linked to such current processes. We also identified that some parts of the scarp are coated by a brown material, that is fine-grained but contains cm-large clasts (Fig. 3e). This brown coating is indurated and fills the irregularities of the bedrock that lies beneath (visible in holes through the coating, Fig. 3e, Supplementary Movie 1). The striae unrelated to sedimentary processes are probably tectoglyphs created by dip-slip fault motion, while the coating resembles fault gouge (or fault breccias) made of rocks crushed during fault slip. We thus interpret the base of the cumulative scarp as a fault mirror, i.e., a surface that has been polished by slip, and with a displacement generally down-dip, with limited lateral motion.

This fault mirror is heterogeneous. It exhibits a light brown color surface over the first ~10 m or more, while above, the scarp is often covered by a dark coating (visible on raw images and 3D models, Fig. 3a–c and S3a, b, c, f), or has a rougher aspect (Fig. S3a, f) and can be incised by small gullies (Fig. S3d, e). Such color and roughness changes are often observed at seismically active faults on land^{38,39}, and are an indicator of surface aging due to weathering of different portions of the fault that have been exhumed at different periods due to seismic displacement. The color changes we observe on the Amorgos fault are also characteristic of active submarine faults^{10,11}, with coatings of Fe–Mn oxide, depositing at very slow rates (1 μm/yr to 0.125 mm/yr⁴⁰) but efficiently darkening surfaces exposed to seawater. Lighter and smoother fault sections have thus been recently exposed.

Locally, the fault mirror is also topped by a thick stripe of fine white sediment (Fig. 3a–c and S3). It is indistinguishable in appearance from the hemipelagic sediment of the wedge below, seen within failure scars (Fig. S1), and can present remnants of stratified sediments from the wedge (Fig. S3d,

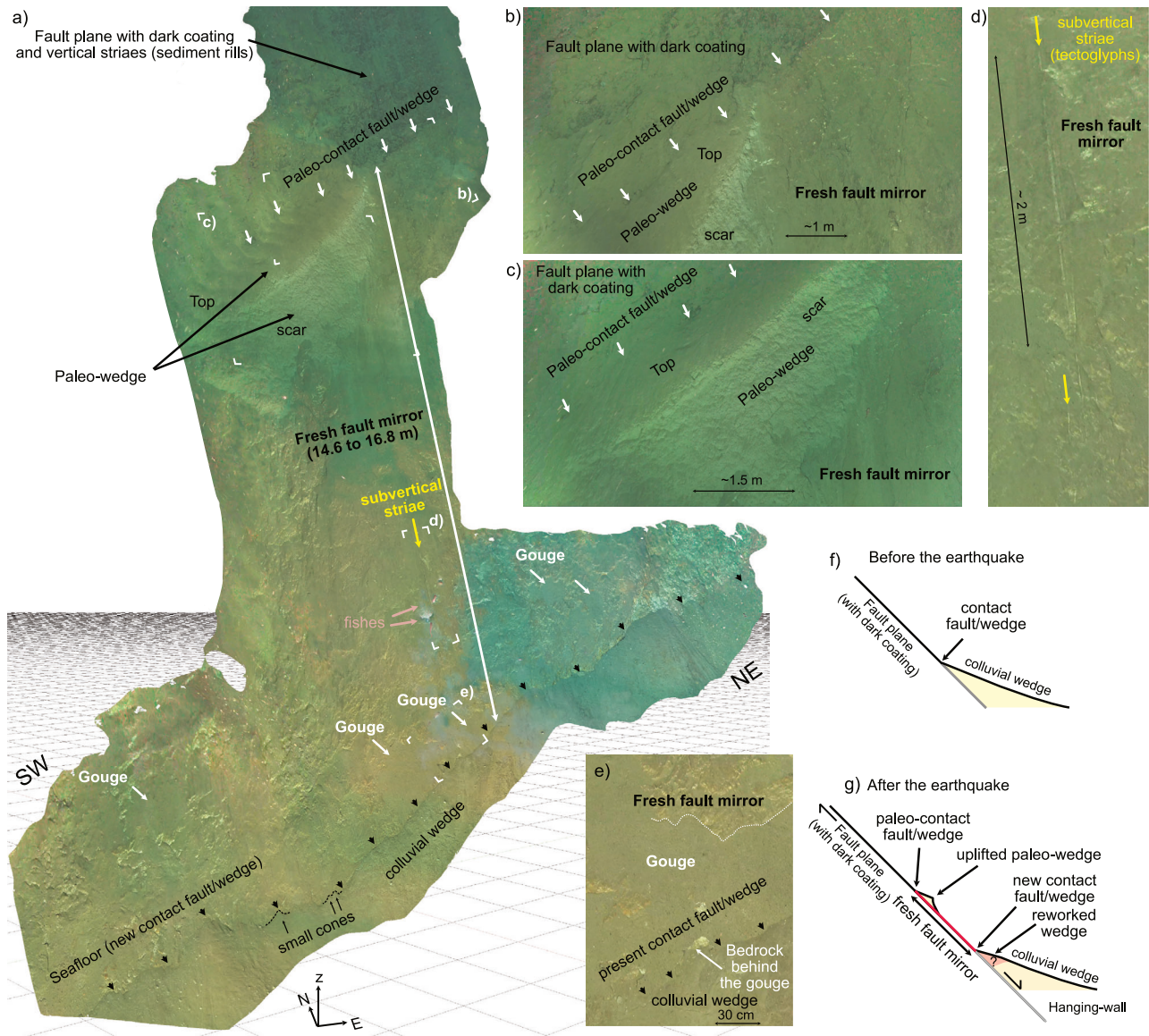


Fig. 3 | The Amorgos fault mirror. **a** 3D textured Digital Outcrop Model and **b–e** raw pictures of the submarine fault mirror observed at the base of the Amorgos fault (location shown on Fig. 2), and associated with the 1956 Amorgos event. It shows striae (**d**), gouge coatings (**e**), and is topped by the remnants of the paleo-wedge portion/section that has been uplifted (**a–c**). We interpret this uplifted paleo-wedge material as a marker of the paleo-contact between the fault and the seafloor

prior to the earthquake that exhumed the fault mirror. In addition, the fresh fault mirror is distinguished from the older fault plane that is darker (due to Mn coating, **a–c**). At this location, the fresh fault mirror is 14.6–16.8 m high (corresponding to the on-fault offset). Supplementary Movie 1, recorded by the HROV, was used to produce this model. The evolution of the fault plane morphology before and after the earthquake is represented by diagrams in (**f**) and (**g**), respectively.

f). We interpret this deposit as the top of the sediment wedge abutting the fault scarp prior to the earthquake, that has been detached from the hanging wall (i.e., the sediment wedge, Fig. 3f, g) by relative down-dip displacement with respect to the footwall (i.e., the fault scarp). Similar tectonically uplifted remnants of soil¹¹ or colluvial wedge¹² are exhumed and uplifted by coseismic fault slip on land.

Georeferenced and scaled digital outcrop models (DOMs) of seven vertical transects have been generated from the HROV 4K videos (Figs. 3 and S3). DOMs allow quantifying the offset of the seafloor by measuring the distance between the present contact of the scarp and colluvial wedge, and the paleo-contact that is now uplifted. This paleo-contact is identified by the top of the paleo-wedge, often associated with a color or a texture change (Figs. 3 and S3). We measured the distance in the direction of the sub-vertical striae, along the fault plane (along-dip). In the seven sites, these on-fault distances range between 9.8 and 16.8 m (Fig. 4 and Supplementary Table S4). We attribute this variability over the ROV path to local

differences in erosion and sedimentation processes (in the form of small failure depressions and small cones, Fig. 3a) that affect the current colluvial wedge since the fault slip. Small-scale variations in the strike, dip, and rake of the fault certainly generate variability in the measured vertical offset⁴³, but not to the extent observed along this ~500 m of fault. Overall, the Amorgos fault exhibits a fresh fault mirror, with a mean height of 12.7 m along-dip, at the location presented here. Similar markers of recent deformation are visible at other places over ~30 km along the fault (Fig. 1), but they are distributed along several fault branches, hindering offset estimation.

Discussion and implications

The preserved remnants of the paleo-wedge and the color difference on the fault scarp argue that the fault exhumation is recent. As no method exists yet that would allow sampling and dating this fault scarp, we must seek seismic events reported in instrumental and historical catalogs^{44–48} in order to discuss its age. Since 1956, no large earthquake has occurred near the Amorgos

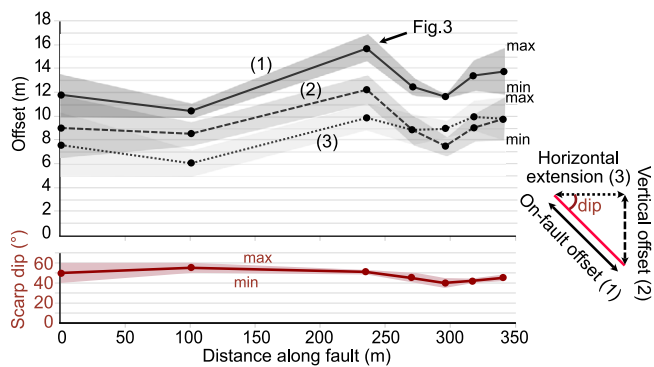


Fig. 4 | Coseismic offset measured along the Amorgos fault. On-fault offsets of the seafloor (1) measured on digital outcrop models (DOMs) of the fault mirror, acquired along several vertical transects, are shown by dots linked by a solid line. The upper bound of the gray surface associated with the line corresponds to the maximum on-fault offsets measured on the DOMs, while the lower bound corresponds with the minimum. The DOM presented in Fig. 3 is located. The fault mirror dip (in red) and its variations (red surface bounded by minimum and maximum dips) were measured in the 1 m AUV DEM, allowing us to calculate the corresponding vertical offsets (2) and horizontal extensions (3) (and their uncertainties shown as gray surfaces). The origin of the x -axis is plotted in Fig. 1a by a black circle on the black HROV path, and DOMs are located on a map in Supplementary Fig. S1.

fault²⁷. Before 1956, a few historical earthquakes, not linked to volcanic activity, struck the island of Amorgos on December 23rd, 1733 and on April 7th, 1891, and were felt with large intensities within the archipelago. Although their locations are not well constrained, the distributions of the macroseismic intensities suggest epicentral areas closer to Sifnos and Chios respectively^{44,45}. Prior to these two events, there are no testimonies indicating that a significant earthquake occurred in the vicinity of Amorgos Island during the past millennium^{44,45,47}, nor before^{44,48}.

Therefore, the recently exposed fault mirror is most plausibly linked to the 1956 event. Tectoglyphs visible on the exhumed fault plane are compatible with focal mechanisms calculated of the mainshock¹³, although they might have formed earlier and deeper. This fault mirror recorded either solely coseismic displacement, or both coseismic and post-seismic slip, with no data available to distinguish one from the other.

Scaling laws provide empirical correlations between earthquake magnitude and rupture parameters, including coseismic displacement, and can be used to evaluate if the displacement observed at Amorgos fault is consistent with the magnitudes of the 1956 event, or exclude it as a possible source. Even though coseismic slip is heterogeneous along a seismic rupture⁴⁹, we can tentatively calculate the seismic moment M_0 and the moment magnitude M_w of an earthquake able to generate the on-fault displacement that is observed, in order to discuss our observations. Different scaling laws link the seismic moment magnitude of an earthquake either to the rupture area A and the mean displacement on the fault⁵⁰, or to the maximum slip at the surface⁵¹.

To use the first formulation⁵⁰, we considered that the earthquake broke the crust and reached the Moho (in agreement with the recently determined seismogenic depth of the Hellenic arc⁵²), along a fault dipping at 60°, with an associated rupture length ranging from 45 km (the main segment) to 75 km (the entire fault system), that the shear modulus is 3.2×10^{10} N/m², and that our observations correspond to the mean coseismic displacement on the fault during the earthquake. These assumptions yield a moment magnitude, M_w , ranging from 8.0 to 8.3 and greater than the magnitude calculated from seismological data¹². If we assume instead that the on-fault offset we observed is representative of the maximum surface displacement along the fault, the second formulation linking magnitude and maximum displacement at the surface⁵¹ yields a moment magnitude M_w of 7.5 (± 0.1), a value in line with the moment magnitude calculated with seismological data¹², which is between 7.2 and 7.8.

Seismic and tsunami hazard implications

The Amorgos tsunamigenic earthquake remains one of the largest normal faulting earthquakes recorded globally and generated the largest tsunami in the Mediterranean Sea for the past two centuries²⁴. Our data provides the first geological evidence that the Amorgos fault was the host of this event. The Amorgos rupture identified and described in this paper shows a vertical offset that varies from 6.4 up to 13.4 m, with a mean of about 9 m at the seafloor (Fig. 4 and Supplementary Table S4, S5). This large vertical displacement, occurring only 1 km from the Amorgos coastline, suggests that the tsunami could plausibly have been triggered solely by the rupture and sudden vertical displacement of the seafloor. This justifies a revision of the role of mass wastings in generating the highest tsunami waves, in this event, at least close to the epicentral area. While mass-wasting seafloor structures are identified in the vicinity (Fig. 1), they are undated, and their link to the 1956 earthquake remains to be established.

Considering the local fault dips, and the on-fault offset, the 1956 earthquake accommodated 9 m of horizontal extension on average (varying between 5 and 11.6 m, Fig. 4, Table S4). A recent study based on GNSS data³¹ quantifies that the fault system between Naxos and Astypalaea must accommodate 4 mm/yr of horizontal extension. This rate implies that about 2250 years of loading is necessary to produce an earthquake similar to the 1956 main shock. If this duration represents the recurrence time for a large earthquake on the Amorgos fault, it may explain the scarcity of events in the 20th century^{45,47} and in historical records^{44,45}, assuming the latter is complete. This possibility should be investigated by future paleoseismological studies in the area, in order to better constrain the seismic and tsunami hazards of the Aegean Sea and Eastern Mediterranean.

The other fault systems in the vicinity of the 1956 epicentral area, namely the Santorini-Amorgos fault, the Anafi-Astypalaea fault system, the Ios fault system, and the Kinairos fault, remain unbroken today (Fig. 1). Although the seismicity of the Santorini-Amorgos region is among the most sustained of the Hellenic Volcanic arc, it is largely triggered by magmatic phenomena²⁸ that are intense around Santorini and other submarine volcanoes (Fig. 1). In historical chronicles, the closest and largest earthquake to have occurred in the vicinity of the fault system was the 42 AD event, located between Santorini and Amorgos, and was probably linked to the volcanic activity of the Santorini volcano, which erupted in 46 AD⁵³. In recent times, only two M_w 6 earthquakes have occurred, in 1911 and 1919, between the Anafi-Astypalaea and Santorini-Amorgos faults⁵⁴, which are both long enough to host earthquakes of $M_w \geq 7$. With the exception of the Santorini-Amorgos fault, the other faults left unbroken show very little microseismicity²⁷. This may suggest that they are either locked and accumulating stress, or that aseismic slip is occurring²⁷. The seismic coupling coefficient (SCC) calculated in this part of the Aegean Sea establishes that the region is strongly coupled (SCC \sim 80%)⁵², except in the vicinity of Santorini (SCC \sim 40%). This indicates that stress is mainly released in the form of large earthquakes in this region and that the Anafi-Astypalaea fault system, the Ios fault system, and the Kinairos fault could break in the future.

Conclusion

Sixty-seven years after the 1956 Amorgos earthquake, we have used submarine vehicles to observe well-preserved and clearly visible traces of a large seafloor rupture along the Amorgos fault, allowing us to identify this as the likely source of the earthquake. This result, together with previous studies of the Les Saintes earthquake^{10,11}, opens the door to submarine explorations looking for the—still debated—fault sources of major historical earthquakes and tsunamis elsewhere, such as the 1783 and 1908 Messina strait events, the 1755 Lisbon earthquake and the 1833 Showa-Sanriku and 1977 Sumba outer-rise events. Comprehensive and detailed submarine geological observations will also lead to a better understanding of the mechanisms behind tsunami triggering⁵⁵. Data from submarine exploration will be particularly important to better predict future sources of underwater earthquakes and tsunamis and to assess how our communities can adapt to these natural hazards, especially where tourism can have a major impact on a region's vulnerability.

Methods

Bathymetry data

During the AMORGOS-22³⁶ and AMORGOS-23³⁷ cruises, we acquired shallow and deep bathymetry using the multibeam echosounder Kongsberg ME70 onboard R/V Europe. Bathymetry was acquired at low speed (2–5 kt) to densify the beams and thus increase the resolution along the steep fault scarp. During the AMORGOS-22³⁶ cruise, we also collected near-bottom, high-resolution bathymetry data using the AUV Idef[®] (IFREMER, France), equipped with a Kongsberg Reson SMF EM2040. The AUV surveyed ~70 m above the seafloor, parallel to the faults. All bathymetric data were processed using GLOBE (IFREMER) and gridded to produce digital elevation models (DEMs) with a 1 m cell size for the AUV and a 10 m cell size for the ship bathymetry (Fig. 2a).

ROV 3D digital outcrop models from video imagery

During the AMORGOS-23³⁷ cruise, we deployed the HROV Ariane (IFREMER) in order to collect video imagery at multiple locations along the fault scarps (Figs. 3 and S3). It is equipped with a 4 K camera (DeepSea Apex SeaCam) mounted on a pan-and-tilt platform at the front of the vehicle, and a second HD camera mounted on the lower-right corner of the vehicle. We surveyed the fault scarp horizontally and vertically, using overlapping tracks at speeds of <0.5 m/s and at distances of ~2–5 m from outcrops. Extracted video frames (every 2 s) were corrected for illumination attenuation prior to processing, using the MATISSE 3D Preprocessing tool⁵⁶ (IFREMER). A structure-from-motion technique was then applied with MATISSE 3D software (v.1.4)⁵⁶ to obtain three-dimensional digital outcrop models that correspond to vertical transects (method A). The HROV navigation data were used as a priori constraints to build the sparse point cloud, and bundle adjustment was performed before densifying the point cloud, meshing, and texturing the models. Terrain models are thus georeferenced and scaled to allow geological interpretations and provide proper scaling for quantitative studies with model resolutions of ~1 cm or better.

To verify the obtained displacements with method A, we used different processing pipelines with two additional methods and algorithms. Method B used the 3DF ZEPHYR software using the HROV navigation data as a priori constraints but without the bundle adjustment. Method C⁵⁷ used the navigation data as a *posteriori* constraint to scale the model. In total, 21 DOMs were obtained that model the seven vertical transects. We used 3DF ZEPHYR's drawing and measuring tools to map the striae, the base, and the top of the fresh fault mirror on each model. We measured the minimum and maximum distances on the model that separate the top and bottom of the fault mirror, in the direction of the striae (i.e., along-dip). Figure S3 presents the seven vertical transects modeled with method A. Table S5 presents the measurements for each of the seven sites, measured with the three methods (A–C). Overall, the on-fault offset measurements show the same trend. Half of the measurements done with method B and C agree with measurements performed with method A within 4.8% and 6.4%, respectively (i.e., median). A few measurements performed with method C depart by up to 26.5% from method A. The DOM of the vertical transect presented in Fig. 3a is very well constrained, as the three methods give similar results (mean on-fault offset of 15.8 ± 0.1). This site displays the largest offset we observed during the dive.

For consistency in the main text and main figures, all on-fault offsets measured and discussed are obtained with MATISSE 3D (method A). Therefore, all vertical offsets and horizontal extensions are derived from method A measurements. Despite this modeling effort, DOMs were mis-oriented due to inaccurate recording of the pan and tilt of the cameras during acquisition. While this does not impact the on-fault distance measurements, we could not use the DOMs to measure the scarp dip and calculate the vertical offset and horizontal extension discussed in the main text. To overcome this limitation, the 1 m DEM derived from AUV surveys allowed us to measure scarp dip, presented in Supplementary Table S4.

Data availability

The 1 m DEM⁵⁸ from the AMORGOS-22 cruise and the 10 m DEM⁵⁹ from the AMORGOS-23 cruise, presented in Fig. 2, are available at <https://doi.org/10.17882/99212> and <https://doi.org/10.17882/90284> respectively.

The HROV-navigation file of dive 1 from the AMORGOS-23³⁷ cruise is available at <https://campagnes.flotteoceanographique.fr/campagnes/18003211/>. Video-derived 3D Digital Outcrop Models⁶⁰ are available at <https://doi.org/10.17882/99228>.

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Author contributions

F. Leclerc designed the study, supervised the ship, AUV and ROV data acquisition and processing, processed the 3D models with Matisse 3D and 3DF Zephyr, interpreted the data and wrote the original paper. S. Palagonia took part in the ship and ROV data acquisition and bathymetric data processing, interpreted the data and revised the paper. N. Feuillet and P. Nomikou took part in the ship, AUV and ROV data acquisition, interpreted the data and revised the paper. D. Lampridou took part in the ship and AUV data acquisition, processed the ship bathymetric data. P. Barrière processed the AUV bathymetric data. A. Dano supervised and processed the AUV and ship bathymetric data. E. Ochoa processed the 3D models with method C and revised the paper. N. Gracias supervised the 3D model processing and revised the paper. J. Escartin designed the study, interpreted the data and revised the paper.

Competing interests

The authors declare no competing interests.

Additional information

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