The 2020 $M_w$ 7.0 Samos (Eastern Aegean Sea) Earthquake: joint source inversion of multitype data, and tsunami modelling

Yu-Sheng Sun¹, Diego Melgar², Angel Ruiz-Angulo, Athanassios Ganas, Tuncay Taymaz³, Brendan Crowell⁴, Xiaohua Xu⁵, Varvara Tsironi, Ilektra Karasante, Seda Yolsal-Çevikbilen, Ceyhun Erman, Tahir Serkan Irmak, Yeşim Çubuk-Sabuncu⁶ and Tuna Eken⁷

¹Department of Earth Sciences, University of Oregon, Eugene, OR 97403, USA. E-mail: yushengs@uoregon.edu
²Institute of Earth Sciences, University of Iceland, 102 Reykjavik, Iceland
³National Observatory of Athens, Institute of Geodynamics, Lofos Nympon, Thission, 11810 Athens, Greece
⁴Department of Geophysical Engineering, The Faculty of Mines, Istanbul Technical University, Maslak 34467, Sariyer, Istanbul, Türkiye
⁵Department of Earth and Space Sciences, University of Washington, Seattle, WA 98195, USA
⁶School of Earth and Space Sciences, University of Science and Technology of China, Hefei, Anhui 230026, P.R. China
⁷Institute for Geophysics, University of Texas at Austin, Austin, TX 78758, USA
⁸Department of Geophysical Engineering, Kocaeli University, 41001 İzmit, Kocaeli, Türkiye
⁹Icelandic Meteorological Office, 105 Reykjavik, Iceland

1 INTRODUCTION

The Aegean Sea and western Anatolia region is one of the most seismically active and most rapidly extending continental regions inside Eurasia because of the complex interactions between the Arabian, African and Eurasian plates (Taymaz et al. 2004, 2007). Due to the collision of Arabia and Eurasia plates, the Anatolian Plate is escaping to west along the North and East Anatolian fault systems (McKenzie 1972; Allmendinger et al. 2007; Barbot & Weiss 2021; Meng et al. 2021; Taymaz et al. 2021). In the northern Aegean Sea, the tectonic structure is dominated by dextral strike-slip faulting, parallel to the North Aegean Trough (Taymaz et al. 1991;

SUMMARY

We present a kinematic slip model and a simulation of the ensuing tsunami for the 2020 $M_w$ 7.0 Néon Karlovásion (Samos, Eastern Aegean Sea) earthquake, generated from a joint inversion of high-rate GNSS, strong ground motion and InSAR data. From the inversion, we find that the source time function has a total duration of ~20 s with three peaks at ~4, 7.5 and 15 s corresponding to the development of three asperities. Most of the slip occurs at the west of the hypocentre and ends at the northwest down-dip edge. The peak slip is ~3.3 m, and the inverted rake angles indicate predominantly normal faulting motion. Compared with previous studies, these slip patterns have essentially similar asperity location, rupture dimension and anticorrelation with aftershocks. Consistent with our study, most published papers show the source duration of ~20 s with three episodes of increased moment releases. For the ensuing tsunami, the eight available gauge records indicate that the tsunami waves last ~18–30 hr depending on location, and the response period of tsunami is ~10–35 min. The initial waves in the observed records and synthetic simulations show good agreement, which indirectly validates the performance of the inverted slip model. However, the synthetic waveforms struggle to generate long-duration tsunami behaviour in simulations. Our tests suggest that the resolution of the bathymetry may be a potential factor affecting the simulated tsunami duration and amplitude. It should be noted that the maximum wave height in the records may occur after the decay of synthetic wave amplitudes. This implies that the inability to model long-duration tsunamis could result in underestimation in future tsunami hazard assessments.

Key words: Transient deformation; Joint inversion; Tsunamis; Earthquake ground motions; Earthquake source observations.
Ganas et al. 2014; Papadimitriou et al. 2020). Since the early Tertiary period, the African plate has been subducting beneath the Eurasia plate in the southern Aegean Sea. This subduction system is being retrograding under gravity, thus leading to its to roll back movement towards Africa. Meanwhile the Aegean plate advances rapidly to the south overriding the almost stalled African slab, causing N–S extension in parts of upper plate of Anatolia and Aegean Sea (Taymaz et al. 1990; Papazachos et al. 1998; Ganas & Parsons 2009; Yolsal-Çevikbilen & Taymaz 2012; Meng et al. 2021). These complex tectonic interplays create active seismicity, so the Aegean Sea and western Türkiye regions have a history of moderate to large earthquakes (Benetos et al. 2006; Yolsal-Çevikbilen et al. 2014; Kassaras et al. 2020; Papadimitriou et al. 2020; Cotin et al. 2020 and references therein). In addition, due to the oceanic environment, the Aegean Sea region is at significant risk of tsunamis generated by earthquakes. This has been well-documented in history and particularly affected the north-central parts of the Aegean Sea region, such as the 20 March 1389, 13 November 1856, 3 April 1881, 23 July 1949 and 09 July 1956 earthquakes (Galanopoulos 1957, 1960; Ambraesys 1960; Guidoboni & Comastri 2005; Yolsal et al. 2007; Okal et al. 2009; Melis et al. 2020). The 1956 Mw 7.8 Amorgos event is the most well-known event that caused run-up heights up to 25 m (Ambraesys 1960; Papazachos et al. 1985; Okal et al. 2009; Dogan et al. 2021). Another triggered activity of tsunami waves at the coasts of Chios Island and Karaburun-Çeşme regions following the 1881 (Mw 6.5) and 1949 (Mw 6.7) earthquakes was reported by Altunok et al. (2005) where the presence of secondary events (i.e. landslides) associated with these major earthquakes were also proposed. More recently, Mw 6.3 and Mw 6.6 earthquakes occurred in June and July 2017, and latter significantly impacted Kos Port in Kos Island in Greece and the southern coast of Bodrum Peninsula in Türkiye (Dogan et al. 2019; Cordrie et al. 2021, 2021).

In this tectonic setting, on 30 October 2020 11:51 UTC, a Mw 7.0 earthquake occurred in the eastern Aegean Sea near the island of Nisyros, affecting Kos Island in Greece and the mainland of Greece. This event is located near the Nisyros Island, which is a volcanic island located in the south of the western part of the Aegean Sea (Taymaz et al. 1990; Ganas et al. 2021; Taymaz et al. 2022; Fig. 1). The Samos earthquake caused a total of 119 deaths, more than 1000 injuries, and some structural damage and building collapses in both areas (Cetin et al. 2020; Mavroulis et al. 2022). The ensuing tsunami impacted and flooded the southern coastal area of İzmir and Samos Island and affected other eastern Aegean Sea islands (Cetin et al. 2020; Dogan et al. 2021; Kalligeris et al. 2021; Triantafyllou et al. 2021). Dogan et al. (2021) measured and investigated tsunami runup and inundation, focusing on a ~110-km-long stretch of coastline in Turkey and found that the largest tsunami run-up was 3.8 m at an inundation distance of 91 m at Akarca (Seferihisar, İzmir). They also found that most people evacuated after noticing the sea recede, which could be considered a successful self-evacuation and indicate the remarkable level of tsunami awareness of the public. Triantafyllou et al. (2021) surveyed along the coasts of Samos and other Greek islands. In Vathy and Karlovasi (NE and NW Samos Island), the maximum run-ups are approximately 2.0 and 1.7 m at an inundation distance of 97 and 80 m, respectively. For the tsunami wave recording, since no near-field tide gauge records are available, the highest wave amplitude is ~10 cm in Kos Island, located ~115 km from epicentre (Heidarzadeh et al. 2021; Triantafyllou et al. 2021). This event also demonstrates once more the complexity of providing tsunami warning in the near field (Cetin et al. 2020).

After the recent series of tsunami (2017 Lesvos, 2017 Kos-Bodrum and 2020 Samos) it seems pertinent to re-assess and update the tsunami hazard and risk in the Eastern Aegean Sea (Triantafyllou et al. 2021). These recent series of tsunamis would also benefit the performance of tsunami warnings and hazard assessment in the region, such as from NOA’s and KOERI’s Tsunami Warning Systems (Papadopoulos et al. 2016; Necmioglu et al. 2021) and the NEAM Tsunami Hazard Model 2018 (Basil et al. 2021). Through the real events, the system and performance could be practised and improved. In the case of Samos, the uncertainty of the initial earthquake parameters could cause a dramatic difference in the hazard regions and levels (Necmioglu et al. 2021). In addition, from the perspective of tsunami hazard assessment, the seismicity would be updated by adding recent earthquakes. The potential fault models of tsunami scenarios could be improved with more reasonable assumptions through inversion works. This would help to further explore the possibility of the variability associated with the rupture area and slip, which affects the probability distribution of tsunami wave heights (Sun et al. 2018). In addition, the conditional probability of tsunami wave height could be better controlled with actual tsunami records.

The U.S. Geological Survey (USGS) generated two possible finite-fault models for the event from an analysis of teleseismic data. However, based on the regional tectonic setting (Cetin et al. 2020), previous point-source and finite-fault source slip inversion studies (Chousianitis & Konca 2021; Ganas et al. 2021; Taymaz et al. 2022) and aftershock distribution (Kiratzi et al. 2021) of the Samos earthquake, the event is more consistent with a north-dipping normal faulting earthquake (Taymaz et al. 1990; Yolsal-Çevikbilen & Taymaz 2012; Meng et al. 2021). To further understand the rupture process and source characteristics of the Samos earthquake, in this study we present a joint kinematic slip inversion of seismic and geodetic data. The Greek and Turkish geophysical real-time monitoring networks provide Global Navigation Satellite System (GNSS) and ground motion data for this study. We also take advantage of interferometric synthetic aperture radar (InSAR) data from the Sentinel-1 satellites that measure the regional deformation along the satellites’ line-of-sight (LOS). From these geophysical data sets, we jointly invert for a kinematic rupture model of the 2020 Samos earthquake.

Previous studies have already focused on the source. Kiratzi et al. (2021) and Taymaz et al. (2022) produced inversions from seismological observations, and Chousianitis & Konca (2021) also inverted for slip jointly using seismic and static and high-rate GPS data. Picka et al. (2022) utilized displacement waveforms acquired from local strong motion stations and static GNSS data to infer the kinematic rupture process. Sakkas (2021) characterized the earthquake source by modellings of combined GNSS and InSAR data. Our contribution here is to include the regional high-rate GNSS, seismic and InSAR data for the first time. We further compare our model with the rupture models from the previous studies in the discussion section.

Regarding the tsunami aspects of this event, the Aegean Sea, being a partially enclosed ocean environment, has the potential to trap tsunami energy, resulting in longer durations compared to more open ocean environments. On the other hand, it may be challenging to equally compare previous global events due to the differences in their source distributions and ocean environments. For example, the duration of tsunami from the 2011 Tohoku earthquake (Mw 9.0) and the 2004 Sumatra tsunami (Mw 9.3) could reach ~6 d (Heidarzadeh & Satake 2014) and ~3 d (Titov et al. 2005), respectively. However, it is reasonable to assume that as the event magnitude increases, there will be a proportional increase in the duration of the tsunami wave. Indeed, Heidarzadeh et al. (2021) observed that the tsunami oscillations are visible above the background noise level for over
24 hr at locations, such as Bodrum and Kos. It can be inferred that the Samos earthquake induced a long-lived tsunami oscillation. To understand the behaviour of this tsunami, we analyse the associated tide gauge records and compare them with tsunami simulations. We show the long duration tsunami from trapping of wave energy in the Aegean Sea by frequency-time analysis and discuss the implication of the resolution of the digital elevation model (DEM) used in tsunami simulations, especially for a long duration tsunami simulation.

2 DATA AND METHODS

We use local geophysical network observations (GNSS and strong ground motion) and InSAR images to model the kinematic rupture pattern of the Samos earthquake. The available gauge records in the Aegean Sea are also involved in this study to analyse the tsunami characteristics. These records are further used to verify the tsunami simulation results and the derived slip model of this earthquake. Here we detail the data processing, the inversion procedure and modelling methodologies.

2.1 Earthquake data

Geophysical data sets of the joint inversion consist of three-component time-series at five high-rate (1 Hz) GNSS stations, five ground motion stations (Fig. 1b) and the ascending and descending InSAR images (Fig. 2). First, for the GNSS data set, the five high-rate (1 Hz) GNSS stations are located around the epicentre and deployed by TUSAGA (Turkish real time kinematic GPS network) and URANUS (Tree-Company network in Greece) (Table S1). The raw GNSS data were processed by using GipysX software and final Jet Propulsion Laboratory orbits (Bertiger et al. 2020). Both solid Earth and ocean tides were removed from the displacements. Then, considering the seismic wave propagation, the time-series of the GNSS-measured displacement were trimmed at 30–60 s after the earthquake origin time (OT) and used in the inverse model (Melgar et al. 2020) (Fig. 3). Secondly, the five strong motion stations are operated by KOERI (Kandilli Observatory and Earthquake Research Institute in Türkiye) and ITSAK (Institute of Engineering Seismology and Earthquake Engineering; the nationwide strong ground motion network in Greece) networks (Table S2). The velocity time-series of integrated acceleration processed by the
Figure 2. Panels (a) and (b) are the ascending and descending InSAR images by phase unwrapping of the interferograms, respectively. Panels (c) and (d) present the resampled InSAR data. The circles are resampled and extracted points used in the inversion. The red square is the SAMU station.

Figure 3. Waveforms of high-rate (1 Hz) GNSS. The grey, black and red lines are the processed displacement, the trimmed waveforms for inversion and the inversion result, respectively. The black and red numbers are the maximum amplitudes of observed and inverted displacement waveforms, respectively.

Engineering Strong Motion Database (ESM Database, Luzi et al. 2020) were trimmed at 50–65 s after the OT and bandpass filtered between 0.2 and 0.5 Hz for the inversion (Melgar et al. 2017; Puglia et al. 2018; Goldberg et al. 2020; Fig. 4). Lastly, we generated InSAR images of this earthquake (Fig. 2). The ascending and descending images of InSAR LOS are produced between acquisitions from 24 and 30 October, and 24 October and 11 November, respectively. The interferograms were generated using an open source InSAR processing software GMTSAR (Sandwell et al. 2011), with back-end processing handled with GMT (Wessel et al. 2019). The data were purely geometrically coregistered (Xu et al. 2017) and filtered with a 200-m low-pass Gaussian filter. To allow radar phase to change properly crossing the narrow straights among the mainland and island, a nearest neighbour interpolation is performed to fill the wet area. Phase unwrapping was then performed using the SNAPHU software (Chen & Zebker 2002) and after unwrapping, we removed a linear ramp estimated using data far away from the coseismic deformation. The LOS vectors were computed pixel-wise to account for the change in look angles due to elevation. Finally, the InSAR images are resampled by a quadtree approach to retain the significant structure of the images (Lohman & Simons 2005; Melgar et al. 2017) for source inversion (Figs 2c and d).

2.2 Inversion

We use the MudPy (Melgar & Bock 2015), an open-source code, and the 1-D regional velocity model (Karakonstantis 2017) (Table 1) for the joint inversion. We consider the GEOFON hypocentre
Figure 4. Waveforms of ground motion (velocity). The grey, black and red lines are the processed velocity record from ESM Database, the trimmed waveforms for inversion and the inversion result, respectively. The black and red numbers are the maximum amplitudes of observed and inverted velocity waveforms, respectively.

Table 1. 1-D velocity model adapted from Karakonstantis (2017).

<table>
<thead>
<tr>
<th>Thickness (km)</th>
<th>$V_p$ (km s$^{-1}$)</th>
<th>$V_s$ (km s$^{-1}$)</th>
<th>Density ($10^3$ kg m$^{-3}$)</th>
<th>$Q_p$</th>
<th>$Q_s$</th>
</tr>
</thead>
<tbody>
<tr>
<td>11.5</td>
<td>5.88</td>
<td>3.40</td>
<td>2.65</td>
<td>600</td>
<td>300</td>
</tr>
<tr>
<td>12.0</td>
<td>5.93</td>
<td>3.43</td>
<td>2.67</td>
<td>600</td>
<td>300</td>
</tr>
<tr>
<td>23.0</td>
<td>7.13</td>
<td>4.10</td>
<td>3.05</td>
<td>600</td>
<td>300</td>
</tr>
<tr>
<td>26.0</td>
<td>7.30</td>
<td>4.20</td>
<td>3.11</td>
<td>600</td>
<td>300</td>
</tr>
<tr>
<td>Infinite</td>
<td>7.90</td>
<td>4.57</td>
<td>3.29</td>
<td>600</td>
<td>300</td>
</tr>
</tbody>
</table>

*From surface to infinite bottom.

(http://geofon.gfz-potsdam.de/eqinfo/event.php?id=gfz2020vimx) as the nucleation point of the rupture. The strike and dip of the fault model are 276° and 48° to the north, respectively. These values approximate the solutions from the GEONF and the USGS focal mechanisms (https://earthquake.usgs.gov/earthquakes/eventpage/us7000c7y0/executive?utm_medium=email&utm_source=ENS&utm_campaign=realtime), and the local fault model from Greek Database of Seismogenic Sources (GreDass, Caputo & Pavlides 2013). The length and width of the fault are set to 102 and 33 km, respectively, and discretized into 3 km by 3 km subfaults (Fig. 1b). The fault model is designed to extend downward from the surface. We assume the rise time is 1.5 s (Somerville et al. 1999). The relative weights between data sets were determined by grid search (Melgar 2014; the Supplementary Text S1). We tested the weight parameters at a time, fixed the remaining variables and decided the value based on the root mean squares (RMSs)

$$||W Gm - W d_{obs}||_2$$  (1)

where $W$ is weighting matrix, and $d_{obs}$ is the values of observation. We found that the weights (1/σ) of horizontal and vertical components of GNSS are reciprocals of 1.6 and 6.4, respectively because of higher noise on vertical component (Melgar et al. 2017). The IKAU station has higher RMSs of waveforms, so we decided to lower its weight to the reciprocal of 3.2 and 12.8. The weights of ground motion and InSAR are reciprocal of 4.0 (Fig. S1). For the rupture speed, we also performed a grid search for each data set and determined the value based on the RMS curves (Melgar et al. 2020). The interval of the minimum misfit RMS trough is 2.7–3.2 km s$^{-1}$, and we selected 3.0 km s$^{-1}$ as the optimum rupture speed for the final inversion model (Fig. S2, the Supplementary Text S1). We use L2-norm (RMS) in the weighting and rupture speed tests, but the RMS shown in result is

$$RMS = \sqrt{\frac{\sum |d_{obs} - d_{inv}|^2}{N}},$$  (2)

where $N$ is number of data, and $d_{inv}$ is the values of inversion result of a certain data set, respectively. The reason is that L2-norm represents the regression degree of the whole inversion matrix but might be mixed and biased by weighting effects. We want to show the inversion results in a more straightforward way, so the RMS of data sets (waveforms and resampled data) are adopted in the result section.

2.3 Tsunami records

The 1-min sampled tsunami records of the eight gauges in the Aegean Sea are provided by the Joint Research Centre (JRC) of the European Commission world sea levels platform (https://webcite.jrc.ec.europa.eu/) and the Intergovernmental Oceanographic Commission (IOC, http://www.ioc-sealevelmonitoring.org/map.php) (Fig. 1a and Table S3). The range of the original wave heights is roughly from −10 to 10 cm (Heidarzadeh et al. 2021). Previous studies (Heidarzadeh et al. 2021; Hu et al. 2022) identified the period of this tsunami source to be ~14.2–23.3 min. The spectra also reveal additional period peaks, such as 4.5 and 6.9 min at Kos, which are attributed to non-source phenomena, such as basin and sub-basin oscillations. These peaks, however, vary at different stations. The lower limit of the period range of the tsunami is generally in range of ~5–10 min (Helal & Mehanna 2008; Toffoli & Bitner-Gregersen 2017). In addition, we consider that the synthetic waveforms may not characterize the behaviour of tsunami waves.
in the shorter periods (Fig. S5 and Text S2). Thus, we adopted the Butterworth filter (Heidarzadeh & Satake 2014) with a bandpass filter boundary of 8–180 min to analyse the recorded and synthetic tsunami waveforms in the time domain. This filter covers the main period and removes the contribution of tides (Simon & Page 2017). However, we only adopted the high-pass filter (180 min) to calculate the spectrograms. This was done to display and compare the spectrum changes of the records and synthetic waveforms over time.

2.4 Tsunami simulation

We model the seafloor deformation resulting from our inferred slip distribution using the elastic dislocation theory (Okada 1985). We assume that the initial perturbation of the water surface is due to the vertical seafloor deformation (Tanioka & Satake 1996; Melgar & Bock 2015; Sun et al. 2018). To model tsunami propagation, we utilize GeoClaw, a finite-volume based suite of solvers for the non-linear shallow water equations using adaptive mesh refinement (AMR) that allows to automatically refine to higher discretization levels (Qin et al. 2019; Clawpack Development Team 2020). Our model was set up to have three different resolutions of the DEM layers, at 30, 15 and 3.75 arcsec. The DEMs of 30 and 15 arcsec are from SRTM15+ (Tozer et al. 2019), and the DEM of 3.75 arcsec is from the European Marine Observation and Data Network (EMODnet, https://portal.emodnet-bathymetry.eu/). Fig. 1 shows the simulation domain, and the finest grids of bathymetry are used at the adjacent sea area of the eight gauges (Fig. S3, the Supplementary Text S2). We simulated 2 d of tsunami propagation. We carried out a bathymetry resolution test that compares the synthetic tsunami waveforms produced by two different DEM layer settings. The first one is mentioned above (main setting, three-layers, 3-L), and the second one uses only the 30 and 15 arcsec resolution (two-layers, 2-L). The results obtained from the different resolutions of the bathymetry are presented in the discussion.

3 RESULTS

3.1 Inversion

Fig. 5 shows the total slip and source time function (STF) from the joint inversion. The estimated moment magnitude of the earthquake is $M_w 7.04\pm4.508 \times 10^{13}$ Nm. There are three asperities on the rupture plane, and most slip occurs on the west side of the hypocentre. The rupture mainly develops westward from the hypocentre, and along strike direction, the dimension of slip is about 72 km towards the west (the total length of fault model is 102 km). For the dip direction, the slip manifests at three different depths: at the beginning of rupture, it develops near the hypocentral depth, then slightly goes updip reaching the upper boundary of the fault, and finally goes downdip ending at the third asperity (Fig. S4). The maximum slip, located next to the hypocentre in the west direction, is $\sim 3.3$ m. The rate vectors show that the rupture is predominantly normal faulting dip-slip motion. The STF presents a total rupture duration of $\sim 20$ s, and there are three main peaks in moment rate function at approximately 4, 7.5 and 15 s corresponding to the development of the three asperities. The result of the STF is almost identical to the moment-rate function of Plicka et al. (2022).

The waveform fitting of GNSS is excellent, and that of ground motion is acceptable (Figs 3 and 4). The RMSs of all waveforms for GNSS and ground motion are 1.36 cm and 0.38 cm s$^{-1}$, respectively. In our model, stations IKAU in GNSS data set and DIDI in the ground motion data set have larger individual RMS for each component. The location relative to the rupture direction may have a greater influence rather than the distance from the epicentre. For example, the IZML, CESM and IKAU have similar distances from the epicentre (∼60 km), but IKAU has higher RMSs. Fig. 6(a) shows the residual distributions of InSAR results. The RMSs of ascending and descending resampled InSAR data are 2.25 and 2.27 cm. Compared with InSAR data, the inverse results fit most observed deformation, but gradually deviate from the observation when the deformation values become smaller and turn into negative (Fig. 6b). We will dwell on this further in the discussion.

3.2 Tsunami

Fig. 7 displays the bandpass filtered tsunami records. The maximum wave heights range from 1.5 cm at Gökçeada to 9.9 cm at Kos. Fig. 8 presents the time-series and spectrograms spanning 4 d of the filtered records. Although the gauge records show tsunami perturbations decayings with time, it is difficult to precisely distinguish the tsunami duration from waveform records themselves. There is ambiguity in determining when the record has decayed in amplitude back to its pre-event noise level. The spectrograms show that the power spectrum density (PSD) increases significantly after the arrival of the tsunami waves, which lasts for several hours. The power then decays and the bandwidth of response period shrinks with time. The response period is approximately from 10 to 35 min (Fig. S6), and for some locations such as Syros, can reach to shorter periods (<5 min) where there may be no significant power before the tsunami arrival (Fig. 8). Through the spectrogram analysis, it is easier to observe that after the earthquake, the power increases significantly and lasts from $\sim 18$ to 30 hr (Fig. S7). The response period and duration vary by location.

The results of the tsunami simulation are filtered in the same way as the real data (Fig. 7). The comparison between the real and synthetic data shows that the first tsunami wave arrivals of the simulation fit well with the observed data, which indirectly proves the accuracy of our slip inversion results. At Syros, however, there is an obvious misfit of the first arrival, and Ren et al. (2022) reported an instrumental malfunction causing the missing of data. Moreover, we found that this misfit of the first period between the synthetic and observed data could also be observed in Heidarzadeh et al. (2021) and Hu et al. (2022). In Fig. 8, the amplitudes and waveforms from the simulation are similar to the records in the first 4–6 hr of the event origin time, with the exception of Heraklion. After the first 4–6 hr, the wave amplitudes in the synthetic data decrease significantly, and the observed waveforms may reach higher or maximum wave heights. Therefore, the simulation would underestimate the observed data, demonstrating the difficulty of producing a long duration behaviour of tsunami propagation. Below we further discuss possible factors affecting this result.

4 DISCUSSION

4.1 Inversion

Our goal is to utilize the waveforms of displacement and strong ground motion and the InSAR to investigate the 2020 Samos earthquake. However, we noticed that the station SAMO (belonging to the HxGN SmartNet network in Greece) has a significant offset (more
Figure 5. (a) Total slip pattern derived from the joint inversion of geodetic and seismic data (GNSS, strong ground motion and InSAR). The black arrows and the yellow star represent the rake direction and the earthquake epicentre, respectively. The circles are aftershocks ($M_w \geq 3$ within 30 d) from Kiratzi et al. (2021). (b) Source–time function (STF) obtained in this study.

Figure 6. (a) The ascending and descending residual of the resampled InSAR data. The red circle represents the resampled InSAR point, the closest to the SAMU station (square). (b) The relationship between observation and inversion. (c) Comparison between the LOS measurements of InSAR images and the GNSS displacement projected onto the satellite look direction.

Figure 7. The observed (black) and synthetic (red) tsunami waveforms. The max number is the maximum positive value of the tsunami wave height.

than 30 cm) and is quite close to the epicentre. Here, we discuss how our model fits the SAMO displacement. We calculate the displacements of the SAMO and other stations based on our slip model with the elastic dislocation theory (Okada 1985). Fig. 9(a) shows the comparison between the GNSS records (Chousianitis & Konca 2021; Ganas et al. 2021) and the simulation. In general, our model fits the vector directions and the amount of offset at most stations. At the SAMO, our model fits the horizontal direction and the value of the vertical component very well, but only fits 53 per cent of the amount of horizontal offset. The SAMU and 093A (belonging to the HEPOS state network of Greece), the rest stations on the Samos Island, also fit well in the horizontal direction, but they exhibit larger
vertical and smaller horizontal displacements, respectively. Fig. 9(b) shows the inversion results without InSAR data. In the horizontal component, the amounts of the SAMO and 093A fit better, but there are slight deviations in the vector direction. The rest of the stations are still similar to the observation. In the vertical component, the value at the SAMO changes from less to greater than the observed value. The value of SAMU is still larger than the observation, but it changes from 1.5 to 2.3 times. In our case, the incorporation of InSAR data serves to rectify and confine the orientation of the horizontal vectors and the magnitude of the vertical components. However, this may compromise the accuracy of values within the horizontal component.

To understand the trend of InSAR misfit in the inversion, we compared the GNSS and InSAR measurements at this area. The GNSS offset of the SAMU station is projected onto the satellite look direction and compared with the resampled InSAR LOS measurements at the closest points to this station (Fig. 6). In Fig. 6(c), the descending InSAR LOS is consistent with the GNSS value, but the ascending datum and GNSS value are inconsistent. From the inversion results, however, the value of ascending datum becomes positive, aligning with the observed GNSS value, and the descending and ascending results are consistent with the GNSS values. Therefore, the trend of the InSAR values in Fig. 6(b) is possibly due to the inconsistent observations and influenced by the higher weighting of the GNSS data in the joint inversion process. This comparison is based on the one of sampling points (sampling with a quadtree approach to retain the significant structure of the InSAR images), which is the closest to the SAMU, so the point and the GNSS station may not be exactly at the same position and may have different offset values. The distance to the SAMU is 2.65 km for the ascending point and 0.29 km for the descending one. Due to this greater distance of the ascending point, it may have a larger difference in value with the SAMU. Furthermore, the previous section demonstrates the effect of the inversion results from the InSAR data. On the other hand, although InSAR measurements are subject to centimetre-level atmospheric perturbation, even after applying some mitigation approach like GACOS (Yu et al. 2017), a possible additional contribution is that the InSAR data were acquired days after the event where extra post-seismic deformation (Goldberg et al. 2020; Ganas et al. 2021; Cirmik et al. 2024) biased the InSAR measurements.

We compare our results with previous studies with various data, including seismic data from Kiratzi et al. (2021) and Taymaz et al. (2022), seismic and geodetic (GPS static offsets and high-rate GPS) data from Chousianitis & Konca (2021) and seismic and static
GNSS data from Plicka et al. (2022) (Fig. 10). Each study used different earthquake parameters, fault plane assumptions, geophysical data sets and inversion packages, but these slip models have some essentially similar features. In Taymaz et al. (2022), Model B and Model C obtained from the joint inversion of teleseismic observations with strong ground motion data are proposed as slip model candidates because the near-field data improves the resolution of rupture details. They select Model C as the final model, based on the misfit and consistency with high frequency back-projected rupture images. With a qualitative comparison of cumulative slip distribution, however, the slip distribution of Model B shows two clear asperities located at the similar areas of our slip model. Therefore, it may be more similar to ours. The results of Chousianitis & Konca (2021), Kiratzi et al. (2021) and Plicka et al. (2022) report a large asperity that overlaps with two asperities in our model. Besides, some studies propose a bilateral rupture process (Lentas et al. 2022; Plicka et al. 2022; Taymaz et al. 2022). This feature is also shown in our snapshot within the first ~8 s (Fig. S4). In general, these slip models from different studies show that the main asperity is located to the west of the epicentre and that the rupture direction is predominantly westward. In addition, the relocated aftershocks ($M_w \geq 3$ within 30 d) from Kiratzi et al. (2021) have good anticorrelation with spatial distribution of these slip distributions (Figs 5a and 10), which is consistent with the observations of moderate and major earthquakes (Melgar et al. 2017, 2020; Chousianitis & Konca 2021 and references therein). In other words, there is low aftershock activity displayed at asperity areas. The analysis of Coulomb stress changes and aftershock activity response in Kiratzi et al. (2021) and Chousianitis & Konca (2021) further shows that the locations of most aftershocks correlate to positive stress changes that increase loading to failure. This supports this anticorrelation and proves the general slip pattern of the Samos earthquake.

We further compared the effective source dimensions (Mai & Beroza 2000; Thingbaijam & Mai 2016) of these slip distributions against the source dimension scaling laws of Thingbaijam et al. (2017) (Fig. 11a). Upon closer inspection, we found that the effective lengths and widths are mostly within the 68 per cent confidence interval and close to the means of the scaling laws. Compared to other studies, the effective length in our fault model is somewhat longer, but still in an empirically reasonable range. This indicates that the Samos earthquake is a mean expected earthquake of rupture dimension. In addition, the peak slips vary between studies. The maximum slips in Taymaz et al. (2022) are 2.7, 3.12 and 3.2 m, respectively; the model of Kiratzi et al. (2021) has peak slip amplitude of ~3.5 m; the maximum slip amplitude in Chousianitis & Konca (2021) reaches 4.6 m; the maximum slip in Plicka et al. (2022) is ~2.4 m; our model is 3.29 m. We should be aware of that the inverted maximum slip would be related to the rupture area (Plicka et al. 2022) and that the measurement of effective dimension is based on the autocorrelation function. Therefore, a model like ours may include some insignificant slip areas. Although their rupture areas and peak slips are different, the models still basically follow the trend of smaller regions having larger slips (Fig. 11a). On the other hand, this would be also related to the strength of the smoothing factor adopted in each study (Plicka et al. 2022). The application of inversion still raises the issue of non-unique solutions and a trade-off with regularization.

For the STFs, in Taymaz et al. (2022), the durations of the optimized point-source mechanism and back-projected analysis are 22 and ~20 s, respectively (see their figs 3a and 8). The durations of Model A and C are ~26 s (Fig. 11b). Kiratzi et al. (2021) reported a duration of ~16 s and identified three main peaks in the STF. Plicka et al. (2022) presented the Samos earthquake using the multiple point source (MPS, Isola NNLS) modelling, which identified three
Figure 10. Comparison with previous studies: Taymaz et al. (2022) (Model A, B and C), Chousianitis & Konca (2021), Kiratzi et al. (2021) and Plicka et al. (2022) (model: LinSlipInv). Our inversion result is shown by the six contour lines (slip = 0.5–3 m with 0.5 m interval), which are enclosed within the fault boundary, the grey dot rectangle. The black dots are relocated aftershocks (MW ≥ 3 within 30 d; Kiratzi et al. 2021). The yellow star is the epicentre used in this study; the white star is the epicentre used in other studies.

Figure 11. (a) The effective source dimension compared against the scaling laws. The panel shows the effective source area versus maximum slip. The dot and dash line represent the intervals of 1 and 2 standard deviations, respectively. Panel (b) the source time functions of the Samos earthquake.

Subevents and provided ~20 s of the total source duration. The finite-fault kinematic rupture model (LinSlipInv, the slip distribution shown in Fig. 10) presents a duration of ~20 s as well. The STF indicates that it experiences three intense moment releases, even though the cumulative slip result includes only one large asperity (fig. 5b in Plicka et al. 2022). We also referred to the solutions from the USGS and GEOSCOPE Observatory (http://geoscope.ipgp.fr/index.php/en/catalog/earthquake-description?seis=us7000c7y0). Both models depict three peaks within the durations of ~20 s. Our model presents a duration of ~20 s and three intense moment releases, which is
consistent with the source characteristics as indicated by these results. Although our model shares similarities with previous results, the results in Chousianitis & Konca (2021) and Model B of Taymaz et al. (2022) significantly differ from the aforementioned numbers. Their long duration tails would reach approximately 32 and 45 s, respectively. Although both inversions allow various rupture velocities that could extend the duration, we note that these models indicated that the rupture rapidly slows down after the main asperity and toward the northwest downdip asperity [see figs 1b and S10 in Chousianitis & Konca (2021), and figs 5a and b in Taymaz et al. (2022)]. Nevertheless, the majority of the seismic moment is released within the first 20 s (Chousianitis & Konca 2021). In summary, although there are two outliers, most studies based on various methods and geophysical data suggest a source duration of ~20 s. From the STFs, the evolution of the released moment generally involves three distinct sections.

4.2 Tsunami

The different filters and processes will affect the measurement of tsunami characteristics. For the waveform measurement, we utilize a bandpass filter; however, it is important to note that the parameters of a bandpass filter may vary from those utilized in other studies, or alternative methods may be used. For example, Heidarzadeh et al. (2021) used a tidal analysis toolbox, TIDALFIT (Grinsted 2008), for removing tidal signals. The resulting maximum wave heights were 11.9 cm at Kos and 5.1 cm at Bodrum, respectively. Then, Hu et al. (2022) applied a high-pass filter to eliminate low-frequency tidal components, yielding de-tided results of 17 and 4 cm at these two stations. Moreover, Dogan et al. (2021) used a fast Fourier transform (FFT) bandpass filter to remove the astronomical tide, obtaining the highest amplitudes of 12 and 3 cm at Kos and Bodrum, respectively. They each present distinct values for the wave heights at the same stations.

We note that in the 4-d-long spectrograms from the tide gauges (Fig. 8), the power of ambient or background waves exist throughout the entire time and overlaps the response period of tsunami waves. These ambient waves may come from atmospheric pressure gradients and seiches in the region (Toffoli & Bitner-Gregersen 2017). The perturbations remain in the ‘tsunami’ period band despite filtering the records, affecting the comparison between real and synthetic data and causing difficulty in more precisely determining tsunami duration.

Figs 7 and 8 indicate that the model fails to reproduce a long duration of tsunami oscillation. The higher wave amplitudes in the records may come after the amplitudes of synthetic waveforms decrease, as observed in locations such as Kos and NOA-03 (Fig. 8). This limitation in the ability of tsunami simulation is critical, as it could impact the accuracy of tsunami hazard assessment. Thus, we hypothesize that bathymetry resolution is one of the factors that could lead to difficulties in modelling the long durations. We conducted a bathymetry resolution test with the two different layer settings, as detailed in Section 2.4. The results in the time domain indicate that the first tsunami arrivals of both models are essentially identical (Fig. 12). For the subsequent waves, the Gökçeada and IDSL-41 gauges are nearly identical, while at Heraklion, they exhibit slight differences. Compared to 2-L, the wave amplitude of 3-L increases at the gauges of Kos Marina, Kos and NOA-03, especially during the first 2–4 hr. After this time, the Kos Marina and Kos still have larger amplitudes, but the 2-L at the NOA-03 has higher wave heights. This observation is evident in the spectrogram (Fig. 13). The PSDs at the Kos Marina and Kos are more intense and wider to respond to the shorter periods, even after 15 hr from the origin time. Although the PSD of 3-L at the NOA-03 is stronger in the first 2–4 hr and slightly responds to the shorter periods in the earlier hours, the intensity of the 2-L PSD lasts longer, until the next day. At the Bodrum gauge, the 3-L produces higher amplitude after the first arrival. In the spectrum analysis, the power of 3-L persists for a longer duration but appears to be more redundant in the first 12 hr. This issue is also evident in the Syros gauge, but the 3-L still covers a broader range of periods with a longer duration. In the time-series, the 3-L at the Syros exhibits significantly higher wave amplitudes, but both demonstrate a clear macroscopic response to the waveform changes, particularly in the amplitude variation from 2 to 4 hr after the origin time.

The resolution of bathymetry is crucial for tsunami simulations because the fundamental equations of tsunami simulation are based on the water depth. A relatively similar study by Felix et al. (2023)
examined the first 15 min of tsunami propagation at a distance less than 8 km from the tsunami source. Their results suggest that the higher resolution results in more accurate arrival times and reduces underestimation of wave heights. Based on our test results, the bathymetry resolution could significantly impact on the complex propagation of tsunamis. A finer bathymetry may improve the numerical model’s ability to simulate tsunamis with longer durations and cover a broader range of response periods. We should also be aware that the model may not accurately capture the behaviour of long duration or predict higher wave heights. Of course, after the first arrival, the complexity of tsunami propagation and oscillation increases due to the involvement of more interactions between tsunami waves and the terrain. However, a more detailed bathymetry would provide a more accurate representation of the seafloor at shallow depths and coastal topography, ideally resulting in a more realistic simulation of tsunami behaviour. Moreover, a
A C K N O W L E D G M E N T S
This work was partially funded by NASA grants 80NSSC19K0360 and 80NSSC19K1104. Part of this research was sponsored by the NASA Earth Surface and Interior focus area and performed at the Jet Propulsion Laboratory, California Institute of Technology. We thank the European Space Agency for access to the Sentinel-1 data. TUSAGA (Turkish real time kinematic GPS network in Türkiye), URANUS (Tree-Company network in Greece), HxGN SmartNet (METRICA company network in Greece), HEPoS (The Greek Cadastre GNSS network), KOERI (Kandilli Observatory and Earthquake Research Institute in Istanbul, Türkiye) and IRTSAK (Institute of Engineering Seismology and Earthquake Engineering nationwide strong motion network in Greece) networks and Engineering Strong Motion Database (ESM Database) data products; for the station maintenance, data acquisition, distribution and processing on-duty watch personnel. We extend our gratitude as well to the Joint Research Centre (JRC) of the European Commission world sea levels platform and the Intergovernmental Oceanographic Commission (IOC) for access to the tide gauge data. The version of GipsyX used to process the raw GNSS data is licensed to BWC at University of Washington. AG was funded by ‘HELPOS-Hellenic Plate Observing System’ (MIS 5002697), which is implemented under the Action ‘Reinforcement of the Research and Innovation Infrastructure’, funded by the Operational Programme ‘Competitiveness, Entrepreneurship and Innovation’ (NSRF 2014–2020) and co-financed by Greece and the European Union (European Regional Development Fund). TT, SYC, CE and TE thank Istanbul Technical University-Research Fund (ITÜ-BAP), the National Scientific and Technological Research Council of Türkiye (TÜBITAK), Turkish Academy of Sciences (TÜBA) in the framework for Young Scientist Award Program (TÜBA-GEHIP), the Science Academy Chamber-Türkiye (BA-BAGEP) and the Alexander von Humboldt Foundation Research Fellowship Award of Germany for financial support and for further providing computing facilities and other relevant computational resources through Humboldt-Stiftung Follow-Up Programme. We have further benefited from fruitful discussions with Oğuz C. Celik, Ercan Yüksel, Coşkun Sari, Nurettin Kaymakçı and Ökmen Sümer for interpretation of neotectonics features observed in the catastrophic area. We would like to thank Dara E. Goldberg for the discussion on the explanation of USGS solutions. We also would like to thank Anastasia Kiratzi, Konstantinos Chousianitis, Vladimir Plicka and František Gallovič for providing their inversion results and data. Finally, we appreciate Jiri Zahradnik and the anonymous reviewer for their valuable efforts that greatly helped in improving the paper.

S U P P O R T I N G I N F O R M A T I O N
Supplementary data are available at GJI online.

Figure S1. (a) Weighting tests. The circle represents the final weighting results. Panels (b), (c) and (d) are the inversion slip patterns of individual GNSS, ground motion (vel) and InSAR data sets, respectively.

Figure S2. Rupture velocity test.

Figure S3. Geographical representation of the areas with 3.75 arcsec resolution during the tsunami simulation. The red dots are gauges. The red rectangles represent that the areas strictly use the finest layer to do tsunami simulation, and the rest of areas will be automatically adjusted the layers that should be used during the simulation. The resolutions of bathymetry with and without white grid are 15 and 3.75 arcsec, respectively, and they are overlapped.

Figure S4. Snapshots of rupture process. The contour is the total slip distribution.

Figure S5. Power spectrum density (PSD, unfiltered) of the records and synthetic waveforms.

Figure S6. Cumulative PSDs of the first half day after the earthquake.
Figure S7. Cumulative PSDs within the period band of 5–35 min.

Table S1. Information of GNSS data. Distance refers to the distance between the epicentre and station. Time refers to the time interval of trimmed data since the earthquake occurred.

Table S2. Information of ground motion data. Distance refers to the distance between the epicentre and station. Time refers to the time interval of trimmed data since the earthquake occurred.

Table S3. Gauge stations and data source.

Please note: Oxford University Press is not responsible for the content or functionality of any supporting materials supplied by the authors. Any queries (other than missing material) should be directed to the corresponding author for the paper.

DATA AVAILABILITY

The MudPy package can be obtained from https://github.com/dmelgarm/MudPy. Data and models produced in this study can be found at https://doi.org/10.5281/zenodo.8415048. This includes the data of the waveform and InSAR image and station locations used for inversion, the fault geometry, the final best fitting rupture models and the tsunami records and synthetic waveforms. In addition, the inversion results can be found at http://equake-rec.info/SRCMOD/searchmodels/viewmodel/s2020SAMS01SUNX (SRCMOD, Earthquake Source Data Database). The data of strong ground motion can be downloaded from Engineering Strong Motion Database (ESM Database, https://esm-db.eu/#/event/EMSC-20201030_0000082). The solution of the GEOFON hypocentre is http://geofon.gfz-potsdam.de/eqinfo/event.php?id=gf2020vimx and the solution of USGS is https://earthquake.usgs.gov/earthquakes/eventpage/us7000c7y0/executive?utm_medium=email&utm_source=ENS&utm_campaign=realtime. The solution of GEOSCOPE Observatory is from http://geoscope.ipgp.fr/index.php/en/catalog/earthquake-description?seis=us7000c7y0.

The local fault model can be found on the Greek Database of Seismogenic Sources (Gredass.unife.it/gredass/GM/). The local fault model can be downloaded from the Joint Research Centre (JRC) of the European Commission world sea levels platform (https://webrichter.jrc.ec.europa.eu/) and the Intergovernmental Oceanographic Commission (IOC, http://www.ioc-sealevelmonitoring.org/map.php). The bathymetry data can be obtained from SRTM15+: https://topex.ucsd.edu/WWW_html/srtm15_plus.html, and the European Marine Observation and Data Network (EMODnet, https://portal.emodnet-bathymetry.eu/). GeoClaw software for modelling geophysical flows over topography can be found online (http://www.clawpack.org/geoclaw).

FUNDING

This work was partially funded by NASA grants 80NSSC19K0360 and 80NSSC19K1104. Part of this research was sponsored by the NASA Earth Surface and Interior focus area and performed at the Jet Propulsion Laboratory, California Institute of Technology. AG was funded by ‘HELPOS-Hellenic Plate Observing System’ (MIS 5002697) which is implemented under the Action ‘Reinforcement of the Research and Innovation Infrastructure’, funded by the Operational Programme ‘Competitiveness, Entrepreneurship and Innovation’ (NSRF 2014–2020) and co-financed by Greece and the European Union (European Regional Development Fund). TT, SYC, CE and TE were supported for their earthquake source studies by Istanbul Technical University-Research Fund (İTÜ-BAP), the National Scientific and Technological Research Council of Türkiye (TÜBİTAK), Turkish Academy of Sciences (TÜBA) in the framework for Young Scientist Award Program (TÜBA-GBİBP), the Science Academy Chamber–Türkiye (BA-BAGEP) and the Alexander von Humboldt Foundation in Germany through Research Fellowship Award and providing computing facilities and other relevant computational resources by Humboldt-Stiftung Follow-Up Programme.

CONFLICT OF INTEREST

The authors declare that they have no conflict of interest.

REFERENCES


Caputo, R. & Pavlidis, S., 2013. The Greek Database of Seismogenic Sources (GredaSS), version 2.0.0: a compilation of potential seismogenic sources (Mw > 5.5) in the Aegean Region. http://gredass.unife.it/.


© The Author(s) 2024. Published by Oxford University Press on behalf of The Royal Astronomical Society. This is an Open Access article distributed under the terms of the Creative Commons Attribution License (https://creativecommons.org/licenses/by/4.0/), which permits unrestricted reuse, distribution, and reproduction in any medium, provided the original work is properly cited.