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Tsunami hazard assessment in the Ionian Sea due to potential tsunamogenic sources – results from numerical simulations

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Abstract. In spite of the fact that the great majority of seismic tsunami is generated in ocean domains, smaller basins like the Ionian Sea sometimes experience this phenomenon. In this investigation, we study the tsunami hazard associated with the Ionian Sea fault system.

A scenario-based method is used to provide an estimation of the tsunami hazard in this region for the first time. Realistic faulting parameters related to four probable seismic sources, with tsunami potential, are used to model expected coseismic deformation, which is translated directly to the water surface and used as an initial condition for the tsunami propagation.

We calculate tsunami propagation snapshots and mareograms for the four seismic sources in order to estimate the expected values of tsunami maximum amplitudes and arrival times at eleven tourist resorts along the Ionian shorelines.

The results indicate that, from the four examined sources, only one possesses a seismic threat causing wave amplitudes up to 4 m at some tourist resorts along the Ionian shoreline.

1 Introduction

The western part of the Hellenic arc is one of the most seismically active areas in Greece and the entire Mediterranean region (Makropoulos and Burton, 1981; Jackson and McKenzie, 1988; Papazachos and Papazachou, 1997). This area has been repeatedly affected by large magnitude earthquakes (Fig. 1) that have caused severe destruction and human loss.

Large tsunami events require the presence of a thick water layer that can be found only in the oceanic domain, but it can also occur in small basins such as the Ionian Sea where many tsunamis have been reported during historical times.

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The Ionian region has an unexpected economic and tourist growth with an increase in coastal population and the development of large leisure areas during recent years, with many parts of coastal cities being a couple of metres above sea level, making them prospective targets of a large-scale disaster, even if the height of the tsunami wave is moderate. This situation requires urgent solutions for an effective risk management and mitigation plan. For this reason, it is essential to define the tsunami potential of the region and this study presents the first results of such an attempt.

The lack of direct records, however, makes the rigorous estimation of the expected tsunami amplitudes rather difficult, and the analysis of available documents remains, somehow, controversial.

Any attempt to assess a tsunami hazard, based on pure statistical methodologies, will not give reliable results because of data deficiency and because they use relationships linking earthquakes to tsunamis that may not be empirically well grounded. This means that alternative approaches to evaluate a tsunami hazard are called upon (e.g. Tselentis et al., 2006). Most often, a solution to the problem is searched for, in terms of a scenario that considers the largest events known to have hit the area of interest in the past history and to simulate these events through numerical modelling.

This is the approach used in the present investigation, focusing on the tectonic deformation mechanics of the potential tsunamigenic faults and their effect on the tsunami hazard in the region.

2 Tectonic setting and tsunamigenic environment

Western Greece (Ionian Islands and Western Peloponnese) represents one of the most seismotectonically active regions of the Mediterranean. It is part of a region of intense deformation located between two major lithospheric plates, the European plate and the African plate. The African plate is moving northwards relative to Eurasia at a rate of 10 mm/year (DeMets et al., 1990). Its leading edge is being subducted along the Hellenic Trench.

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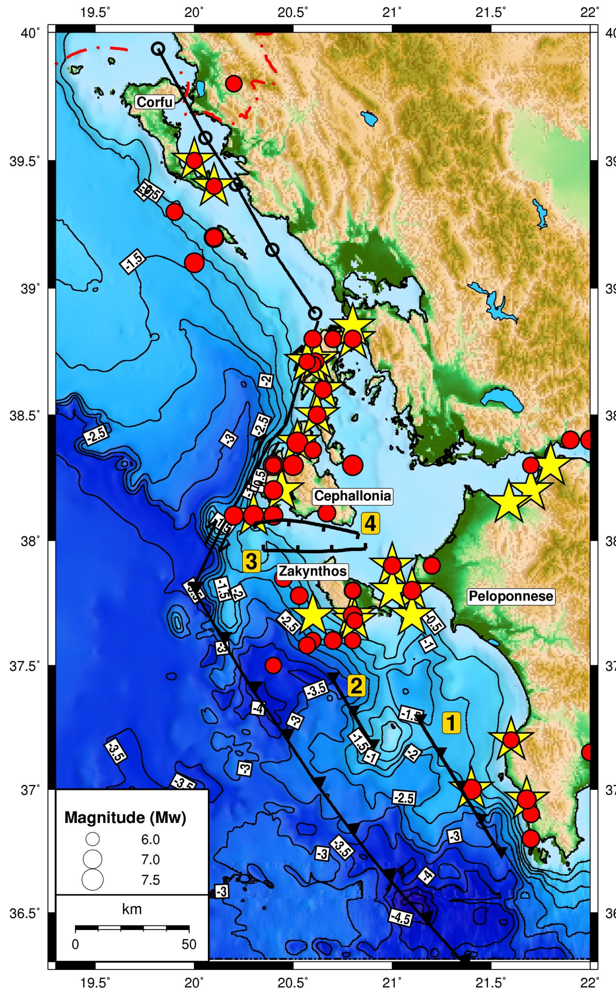

Fig. 1. Earthquakes with $M>6$ occurred along the Ionian Sea and the potential tsunamigenic sources (numbers in yellow squares) considered in the present investigation; contour depths are in km. Stars, denote tsunami reported sides.

Seismological investigations (Jackson, 1994) and geodesy (Noomen et al., 1995) reveal that the relative motion across the Hellenic trench is NE-SW and greater than 5 cm/yr, while GPS measurements (Kahle et al., 1993) show a distinct movement of western Greece to the SW relative to Italy with an average of 120 mm between 1989 and 1993.

Seismotectonics of western Greece has been extensively studied (e.g. Haslinger et al., 1999). The major tectonic features, which set the Ionian Sea as an area of high seismicity are: the subduction of the African plate beneath the Aegean microplate along the western Hellenic trench, and the Cephallonia transfer fault at the northwestern end of the Hellenic arc, which is a major strike-slip fault that links the subduction boundary to the continental collision between the Apulian microplate and the Hellenic foreland.

Fault plane solutions, for several shallow and intermediate depth earthquakes in the Hellenic arc, have been published (Taymaz et al., 1990; Papazachos et al., 1991). Their results indicate that in the western part of the Hellenic arc, the thickness of the shallow seismogenic layer covers the upper 20 km of the crust. According to Papoulia et al. (2008), the area extending from Zakynthos to Messinia is limited by two main fault zones: the one striking NE-SW, offshore Kyllini, dipping to the south east, and the other striking EW, offshore Pylos, dipping to the north. According to Mariolakos et al. (1985) and Papanikolaou et al. (2007), these fault zones are active and are the offshore extensions of EW trending faults of Western Peloponnese.

The western Greece area has repeatedly suffered large offshore earthquakes (i.e. 1886 Philiatra M7.3, 1893 Zakynthos-Keri M6.5, 1899 Kyparissia M6.5, 1947 Pylos M7.0, and 1997 Gargaliani M6.6), which have caused damage and human loss (Papazachos and Papazachou, 1997). The presence of tsunamis have been observed a few times in the past, associated with offshore earthquakes (i.e. 1633 and 1886), affecting near field as well as remote coastal segments (Table 1).

The damage was minor from these events, due to the sparse population of the region in the early years. Today, a large earthquake along the Ionian sea fault system would damage coastal communities, and its effect would be enhanced by sea waves triggered from the seafloor displacement.

Taking into account the Ionian Sea tectonic regime and the tsunamogenic events depicted in Table 1, we have selected four potential seismic sources that could represent a tsunami hazard in the region (Fig. 1). The geographic location, strike, length and other parameters of each source were derived from existing fault maps, available reflection profiles and relevant seismological references and are presented in Table 2.

3 Elastic dislocation modelling

The traditional approach in modelling a tsunami generation is based on solving the hydrodynamic equations with boundary conditions at the seafloor corresponding to a static displacement caused by the earthquake source (e.g. Okal, 1982; Comer, 1984).

Offshore shallow earthquakes result in a seafloor coseismic deformation that can trigger a tsunami. Most studies of earthquake generated tsunamis use Mansinha and Smylie's (1971) formula or Okada's (1985) analytical formula based on elastic dislocation theory to predict seafloor displacement due to an earthquake and to model initial water displacement (e.g. Legg et al., 2004; Tselentis et al., 2006). This approach is being adopted in the present investigation.

In general, we assume that this coseismic deformation is much more rapid than the characteristic time involved in the wave propagation, and that the length scale of the seafloor

Table 1. Possible, locally generated, tsunami along the western Hellenic arc fault system. K_o is the intensity of the wave and it refers to Sieberg's modified intensity scale (1932).

No	Year	Month	Day	Lat/Lon	Region	M	K_o	References
1	1622	May	5	37.60/21.00	Zakynthos Is., Ionian Sea	6.6	?	Ambraseys (1960), Galanopoulos (1960), Antonopoulos (1980)
2	1633	Nov	5	37.70/20.80	Zakynthos Is., Ionian Sea	7.0	2	Ambraseys (1962), Papazachos and Dimitriu (1991), Papazachos et al. (1985), Papazachos et al. (1986), Antonopoulos (1973)
3	1636	Sep	30	38.10/20.30	Kefalonia	7.2	3	Papazachos and Dimitriu (1991), Papazachos et al. (1985), Papazachos et al. (1986), Antonopoulos (1973)
4	1723	Feb	22	38.60/20.65	Lefkada	6.7	2+	Ambraseys (1962), Galanopoulos (1961), Papazachos and Dimitriu (1991), Papazachos et al. (1985)
5	1732	Nov		39.40/20.10	Corfu	6.5	2	Ambraseys (1962), Papazachos and Dimitriu (1991), Papazachos et al. (1985), Papazachos et al. (1986), Antonopoulos (1973)
6	1791	Nov	2	37.90/21 00	Zakynthos Is., Ionian Sea	6.8	3	Ambraseys (1962), Mallet (1850–1858), Papazachos and Dimitriu (1991), Papazachos et al. (1985),
7	1804	Jun	8	38.10/21.70	Patras Gulf	6.4	3	Ambraseys (1962), Hoff (1841), Galanopoulos (1960), Papazachos and Dimitriu (1991)
8	1805	Jan	8	38.30/21.80	Patras Gulf		?	Ambraseys (1962), Hoff (1841), Galanopoulos (1960), Papazachos and Dimitriu (1991)
9	1820	Mar	17	38.80/20.80	Lefkada		?	Soloviev et al. (2000)
10	1820	Dec	29	37.80/21.10	Zakynthos Is., Ionian Sea	6.9	3+	Papazachos and Dimitriu (1991), Papazachos et al. (1985), Papazachos et al. (1986), Antonopoulos (1973)
11	1821	Jan	9	38.15/21.59	Patras Gulf		4	Ambraseys (1962), Mallet (1850–1858), Galanopoulos (1938)
12	1825	Jan	19	38.70/20.60	Ionian Sea	6.8	3	Soloviev et al. (2000), Soloviev et al. (2000)
13	1835	Jul	12	37.70/21.10	Zakynthos Is., Ionian Sea		2	Ambraseys (1962), Antonopoulos (1980)
14	1867	Feb	4	38.39/20.52	Kefalonia	7.4	2	Papazachos and Dimitriu (1991), Papazachos et al. (1985), Papazachos et al. (1986), Antonopoulos (1973)
15	1867	Apr	10	38.20/20.44	Lixouri, Kefalonia		2	Ambraseys (1960), Soloviev et al. (2000)
16	1869	Dec	28	38.85/20.80	Lefkada	6.4	2	Ambraseys (1962), Galanopoulos (1952)
17	1883	Jun	27	39.50/20.00	Ionian Sea		3	Ambraseys (1962), Galanopoulos (1960), Antonopoulos (1980), Soloviev et al. (2000)
18	1886	Aug	27	37.00/21.40	W. Filiatra	7.5	3	Ambraseys (1962), Vidal (1886), Ornstein (1889), Galanopoulos (1960), Galanopoulos (1960b)
19	1889	Jan	22	37.20/21.60	Gulf of Kyparissia	6.6	3	Papazachos and Dimitriu (1991), Papazachos et al. (1985), Papazachos et al. (1986), Antonopoulos (1973)
20	1893	Apr	17	37.68/20.81	Zakynthos Is., Ionian Sea	6.4	2	Ambraseys (1960)
21	1897	Dec		37.80/21.00	Ionian Sea			Antonopoulos (1980)
22	1898	Dec	3	37.80/21.00	Ionian Sea		3	Ambraseys (1962), Antonopoulos (1980), Soloviev et al. (2000)
23	1899	Jan	22	37.20/21.60	Kyparissia	6.5	3	Ambraseys (1962), Mitzopoulos (1900),
24	1914	Nov	27	38.72/20.62	Lefkada	6.3	4	Ambraseys (1962), Galanopoulos (1952), Papazachos and Dimitriu (1991)
25	1915	Aug	7	38.50/20.62	Lefkada	6.7	3	Ambraseys (1962), Galanopoulos (1960), Galanopoulos (1952), Papazachos and Dimitriu (1991)
26	1947	Oct	6	36.96/21.68	W. Pylos, Methoni	7.0	3	Ambraseys (1962), Galanopoulos (1952), Papazachos and Dimitriu (1991)
27	1948	Apr	22	38.71/20.57	Lefkada Vassiliki	6.5	3	Ambraseys (1962), Galanopoulos (1955), Galanopoulos (1960), Papazachos and Dimitriu (1991)

Table 2. Fault parameters of the selected potential tsunamigenic sources used for the numerical simulation (FAUST database, Stucchi et al., 2001; Papazachos and Papazachou, 1997).

Id	Fault Name	Type	Strike (deg)	Dip (deg)	Rake (deg)	Slip (cm)	Length (km)	Width (km)	M_w	Depth (km)
1	Filiatra	thrust	329	30	121	275	70	45	7.5	2
2	Zakynthos	thrust	329	22	120	107	35	22	7	2
3	Cephalonia South	normal	270	60	–90	42	37	13	6.5	2
4	Cephalonia North	normal	110	60	–90	107	40	22	7.1	2

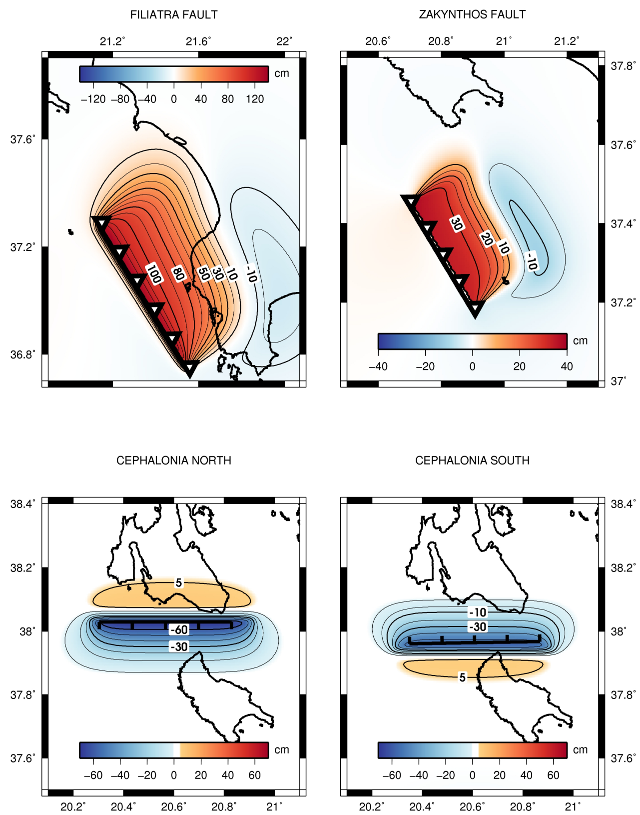

Fig. 2. Sea bottom expected displacements due to the activation of the four sources considering the source parameters depicted in Table 2.

Table 3. Sea bottom deformations due to the activation of the four potential sources.

Id	Fault Name	Vertical Calculated Displacement (cm) (Poly3D)	
		Min (Subsidence)	Max (Uplift)
1	Filiatra	–19	135
2	Zakynthos	–12	39
3	Cephalonia South	–56	7
4	Cephalonia North	–67	8

deformation is much larger than the water depth. This allows us to define the initial subsurface deformation as being equal to the coseismic vertical displacement.

Table 2 presents the source models of the selected tsunamigenic sources used in the present study. The corresponding fault slips u, were determined from (Papazachos and Papazachou, 1997)

$$\log u = 0.82\,M - 3.71 \tag{1}$$

where M is the magnitude.

Initial conditions for tsunami propagation from all four selected potential seismic sources are determined from the static displacement field by using the three-dimensional elastic boundary element method algorithm POLY-3D (Thomas, 1993), assuming a Poisson's ratio of 0.25.

Next, we calculate the sea-bottom strain field for each one of the four fault scenarios considering the source characteristics depicted in Table 2. The obtained results are shown in Fig. 2, and the corresponding vertical displacements are presented in Table 3.

Calculations of seafloor displacements are made assuming a homogeneous elastic structure (Poisson's solid). The effect of elastic inhomogeneity on surface displacements and tsunami waveforms is not included in this study. The near shore tsunami amplitude values predicted in this investigation ignore the possibility of large submarine landslides being triggered by the earthquake, and this study focuses on the tectonic sources only.

4 Tsunami modelling

The deformation field, obtained above, is used as an initial condition of the hydrodynamic computation. This constitutes a legitimate approximation, since the deformation of the seafloor always takes place more rapidly than the tsunami waves can propagate the sea surface deformation away from the source region.

Numerical simulations are useful tools for analysing tsunami propagation and coastal amplification. The tsunami waves, generated by earthquakes, depend on the size and the impact of the source mechanism on the displaced water. Despite the differences in the underlying physics of wave propagation in the sea and solid earth, the tsunami wave field emanating from an earthquake source can be thought of as an extension to the seismic wave field (Okal, 1982).

There are at least three kinds of tsunami propagation and inundation models in common use among tsunami scientists: nonlinear shallow-water models (NSWMs), Boussinesq-type long-wave models, and complete fluid dynamics models, which all stem from the work of Peregrine (1967). NSWMs possess some distinctive advantages that make them very suitable for modelling flows occurring in shallow depths (Brocchini et al., 2001). We perform our study with the nonlinear shallow-water wave model. Since tsunami wavelengths are much larger than the sea depth, tsunamis are considered as shallow-water waves following the long-wave theory.

Long-wave theory is used (where the ratio of water depth to wavelength is small) in which the vertical acceleration of water particles is negligible compared with the gravitational acceleration, and the hydrostatic pressure approximation is used. The nonlinear terms are kept for their use where needed, which is the case in very shallow water (from the tsunami point of view). In addition, we are interested in this investigation on near-field tsunamis, that is, those whose propagation distance is less than 200 km. Therefore, Cartesian coordinates can be used. The vertical integrated governing equations (Dean and Dalrymple, 1984) can be written (after setting the momentum correction factors equal to unity, with Coriolis effect omitted):

$$\frac{\vartheta M}{\vartheta t}+\frac{\vartheta}{\vartheta \chi}\left(\frac{M^2}{D}\right)+\frac{\vartheta}{\vartheta y}\left(\frac{MN}{D}\right)+gD\frac{\vartheta \eta}{\vartheta \chi}+\frac{gk^2}{D^{7/3}}M\sqrt{M^2+N^2}=0 \quad (2)$$

$$\frac{\vartheta N}{\vartheta t}+\frac{\vartheta}{\vartheta \chi}\left(\frac{N^2}{D}\right)+\frac{\vartheta}{\vartheta y}\left(\frac{MN}{D}\right)+gD\frac{\vartheta \eta}{\vartheta \chi}+\frac{gk^2}{D^{7/3}}N\sqrt{M^2+N^2}=0 \quad (3)$$

$$\frac{\vartheta \eta}{\vartheta t}+\frac{\vartheta M}{\vartheta \chi}+\frac{\vartheta N}{\vartheta y}=0, \quad (4)$$

where η is the water surface elevation, t is time, x and y are the horizontal coordinates in zonal and meridional directions, and M and N are the discharge fluxes in the horizontal plane along x and y coordinates given by:

$$M=U\,(h+\eta)=U\,D \quad (5)$$

$$N=V\,(h+\eta)=V\,D \quad (6)$$

where U and V are the vertically averaged horizontal particle velocities and $D=h\,(x,y)+\eta$ is the total water depth, $h\,(x,y)$ is the undisturbed basin depth, g is the gravity acceleration, and k is the bottom friction coefficient. For the simulation reported here and considering the prevailing bottom morphological features, k has been set equal to 0.025 (Fujima, 2001).

For tsunami propagation, the numerical model employed in this study was developed by N. Shuto and F. Imamura, namely of the Numerical Analysis Model for Investigation of Near field Tsunamis version 2 (TUNAMI-N2). This is one of the key tools for the developing studies for propagation and coastal amplification of tsunamis in relation to different initial conditions. It solves nonlinear shallow water equations in Cartesian coordinates using the leap-frog scheme of finite differences. Also, a similar methodology is used in the numerical model of the method of splitting tsunami (MOST) developed by Titov and Synolakis (1998). These are the only two existing nonlinear shallow water codes, validated with laboratory and field data (Yeh et al., 1996).

In this investigation, we use bathymetric data with a grid size of 87 m. The total number of grid points in the computational domain was 8 055 799 consisting of 2741×2939 points. The time step is selected as 0.28 s to satisfy the CFL stability condition (e.g. Tselentis et al., 2006; Goto and Ogawa, 1992). The duration time of wave propagation was 60 min.

Our simulations have two classes of products: we first compute the snapshots of the sea surface displacement all over the grid at time intervals of 4 min. Second, we compute the time series of water surface elevations (mareograms or virtual gauges) at eleven selected locations along the Ionian shorelines possessing increased tsunami hazard due to recent tourist development.

5 Discussion of the results

Figure 3 presents the snapshots of the water surface displacement at 4-min intervals up to 20 min, for Filiatra fault (source 1) and Fig. 5 presents the corresponding mareograms at the 12 most important tourist resorts along the Ionian shoreline (Fig. 4). This potential tsunami source is an offshore fault recognised by a "clustering" of the seismicity; it has a NNW-SSE direction almost parallel to the nearby Peloponnesus coast, and it is a thrust fault with a length L=70 km and a width W=45 km (Papazachos and Papazachou, 1997; Papanikolaou et al., 2007). Several moderate to strong earthquakes have been attributed to this fault; in 1886 this was the source of the M=7.5 event (estimated from the observed maximum intensity X of MM scale), also the 1899 Kyparissia (M=6.5; Papazachos and Papazachou, 1997) and the 1919 Kyparissia (M=6.3; Papazachos and Papazachou, 1997) earthquakes, have been related to this fault.

As we can see from Fig. 3, after 4 min the wave arrives at Pilos city coastal area, where it reaches a maximum amplitude of 1.6 m after 6 min. After 8 min, the wave arrives at Proti island where it reaches the greatest amplitude observed

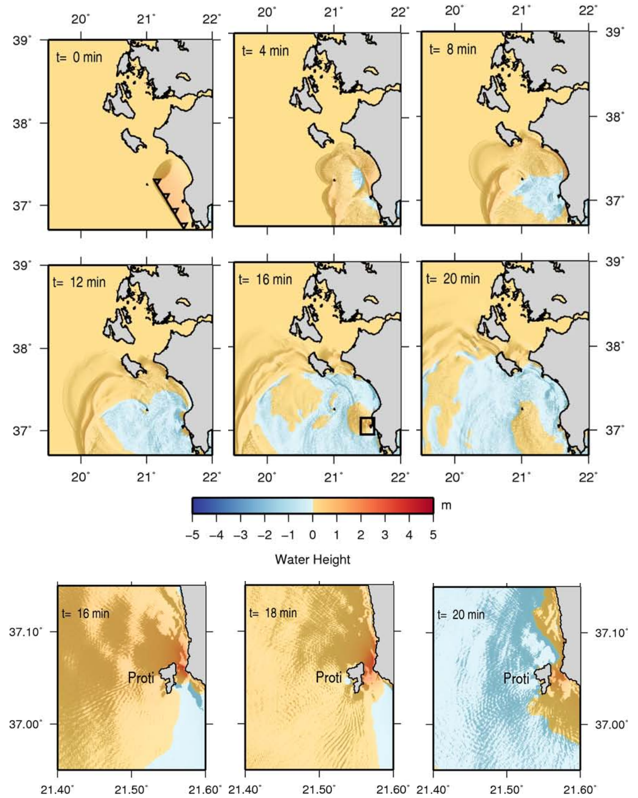

Fig. 3. Snapshots of the tsunami propagation from the source 1 (Filiatra fault) at 4-min intervals. The propagation close to Proti islet is presented in the last row, at a higher magnification.

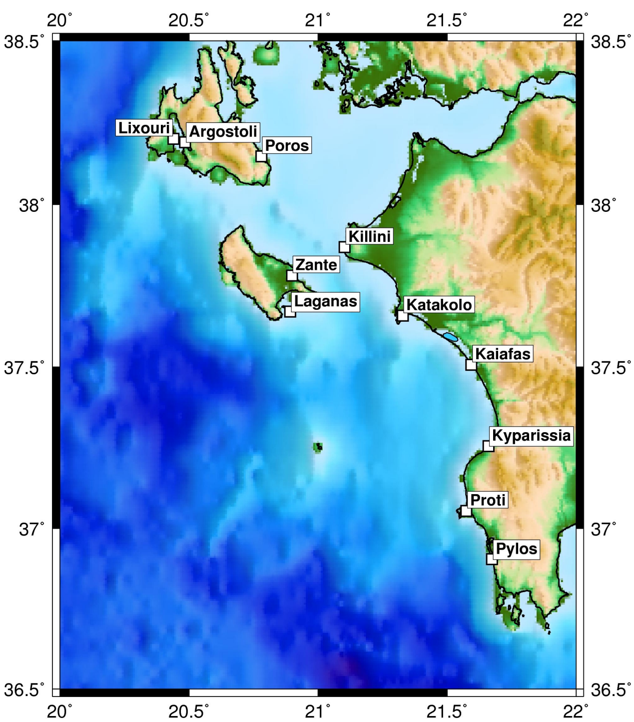

Fig. 4. Locations of the gouges considered in the present investigation, which represent the most populated tourist resorts along the Ionian shoreline.

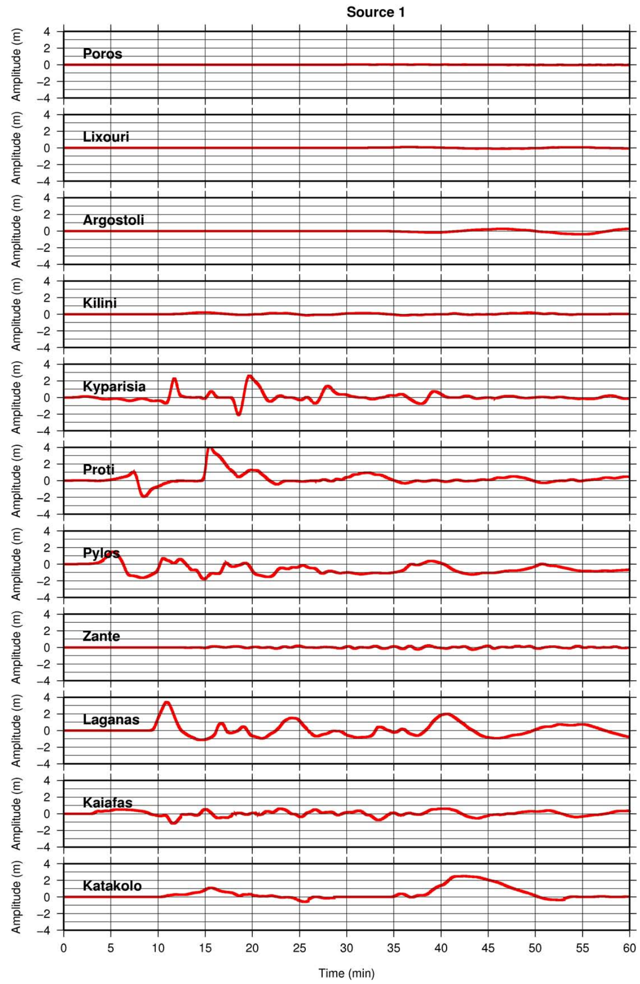

Fig. 5. Calculated mareograms at the locations shown in Fig. 4, for source 1 (Filiatra fault).

in this simulation, 4 m at 16 min. For this reason, we focus in this area depicting tsunami propagation snapshots at a higher magnification.

After 9 min, the wave arrives at the tourist resort of Laganas bay in Zakynthos, where it reaches a maximum amplitude of 3.5 m after 11 min. This is a heavily populated region during the summer with thousands of tourists in the coastal zone and needs specific attention and tsunami risk reduction planning.

For all the other locations, wave amplitudes are relatively small with the exception of Kyparisia region where 13 min and 19 min after the activation of the fault it reaches an amplitude of 3 m. This is also a heavily populated tourist area during the Summer season.

Figure 6 shows snapshots of the tsunami propagation corresponding to the activation of the Zakynthos fault (source 2). The corresponding mareograms at the locations depicted in Fig. 4 are presented in Fig. 7. This potential source is mapped according to seismological evidence (FAUST

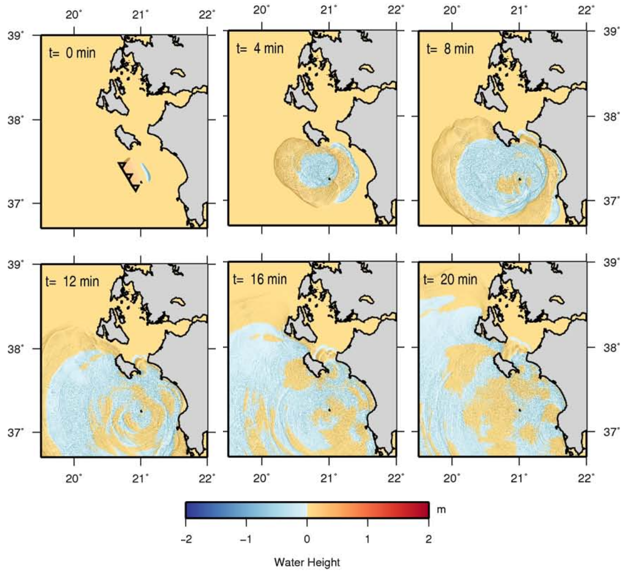

Fig. 6. Snapshots of the tsunami propagation from the source 2 (Zakynthos fault) at 4-min intervals.

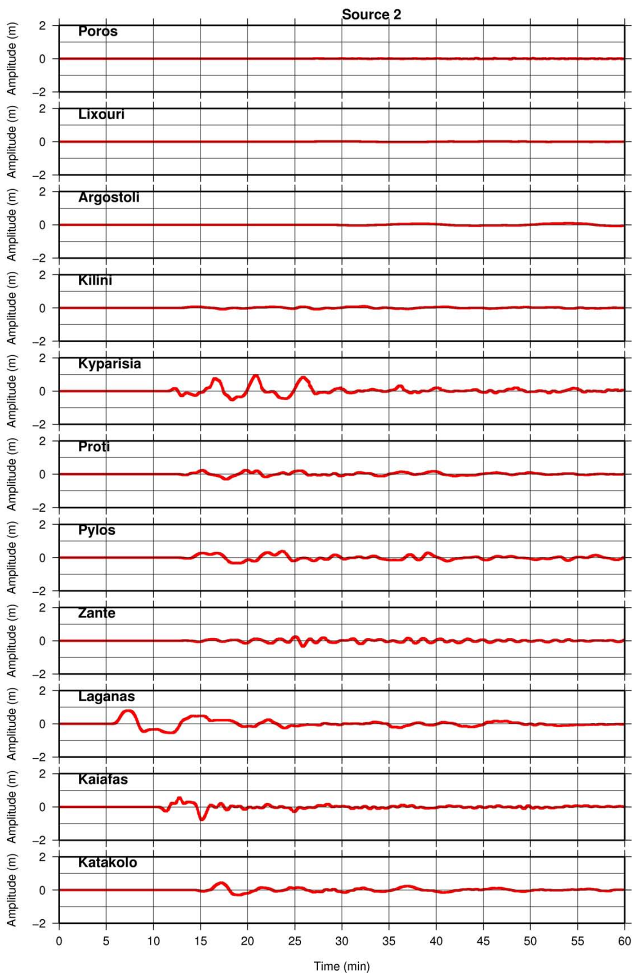

Fig. 7. Calculated mareograms at the locations shown in Fig. 4, for source 2 (Zakynthos fault).

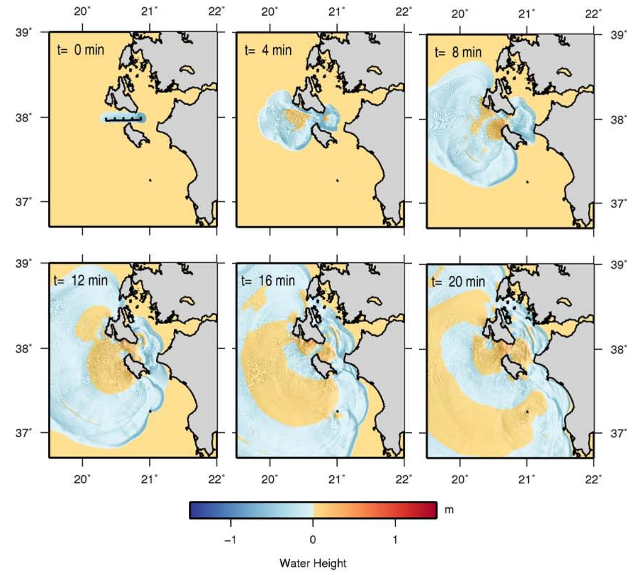

Fig. 8. Snapshots of the tsunami propagation from the source 3 (Cephalonia South fault) at 4-min intervals.

database) proposed by Ambraseys and Jackson (1990) on the basis of the 1958 earthquake with a magnitude M=7.0. It is a thrust fault with a length L=35 km and a width W=22 km. The focal mechanism adopted by the FAUST database is consist with focal mechanisms calculated from recent earthquakes located in the area South of Zakynthos.

Judging from the above figures, we can see that the activation of this particular source has little effect on the tsunami hazard of the region under investigation. The greatest wave amplitudes are reached at Kyparisia bay where they become 1 m at 16, 21 and 25 min after the source activation. Also, at the Lagana bay region, they reach an amplitude of 0.9 m 7 min after the activation of the source.

Figure 8 shows the snapshots of the tsunami propagation corresponding to the activation of the Cephalonia South fault (source 3). The corresponding mareograms at the locations depicted in Fig. 4 are presented in Fig. 9. This source is located in the sea area North of Zakynthos island. According to Monopolis and Bruneton (1982) the northern flank of Zakynthos is steep and is affected by large normal faulting. It is suggested that this normal fault caused the 1554 earthquake, while the maximum expected magnitude from this source is M=6.5 (Stucchi et al., 2001).

In this case, we see that the activation of Cephalonia South source possesses no significant tsunami hazard in the region, since all the wave amplitudes are below 0.5 m. The greatest amplitude is observed at Laganas bay, where it becomes 0.4 m.

Sources 3 and 4 (Cephalonia South and North faults) are mapped in Fig. 1 according to Lagios et al. (2007), where it is mentioned that major faults having the length of several tenths of kilometres in the marine area between Cephalonia

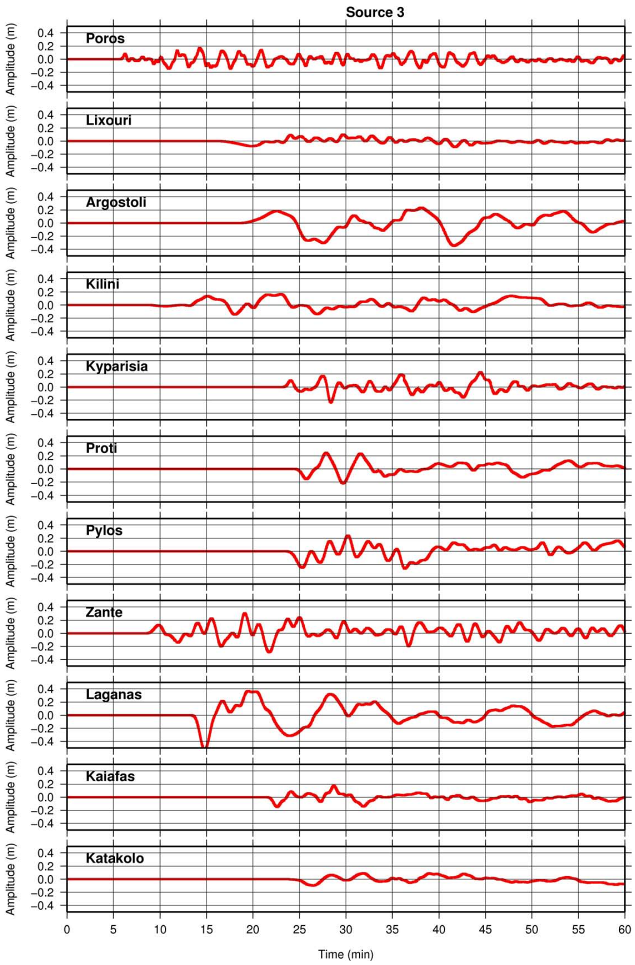

Fig. 9. Calculated mareograms at the locations shown in Fig. 4, for source 3 (Cephalonia South fault).

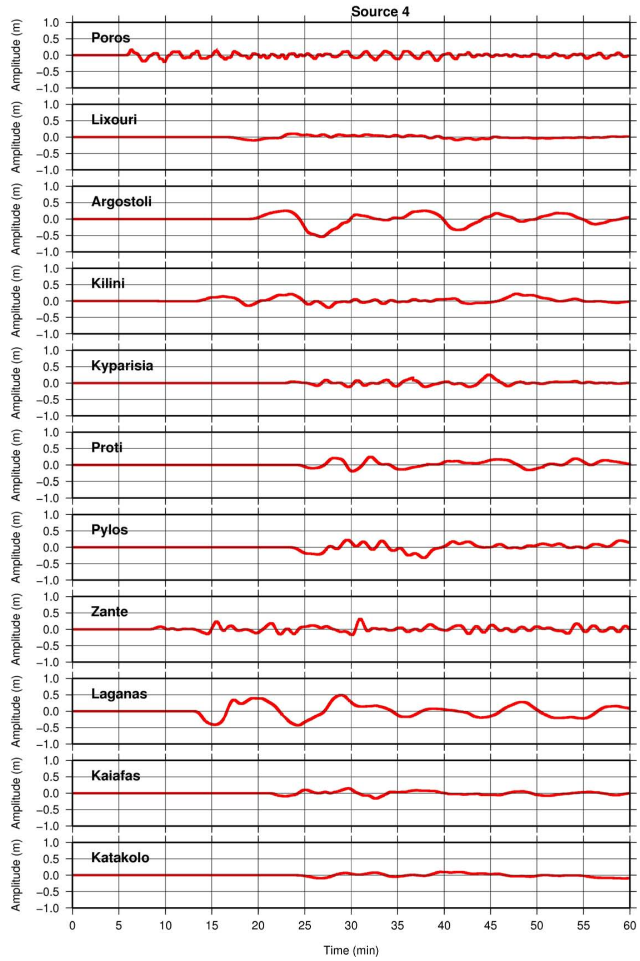

Fig. 11. Calculated mareograms at the locations shown in Fig. 4, for source 4 (Cephalonia North fault).

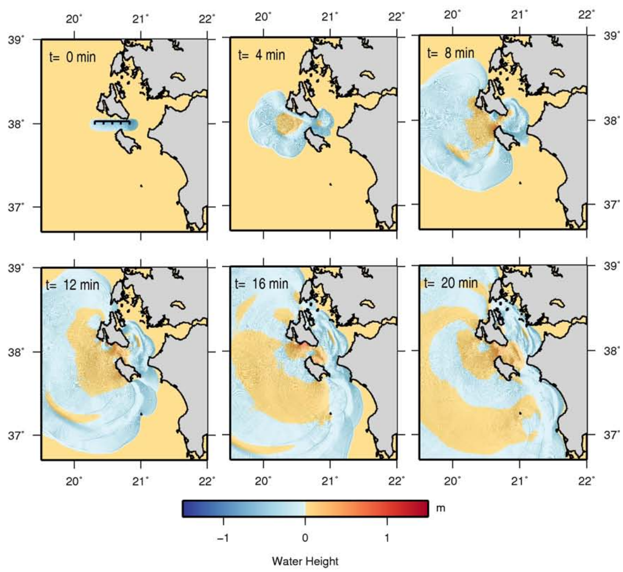

Fig. 10. Snapshots of the tsunami propagation from the source 4 (Cephalonia North fault) at 4-min intervals.

and Zakynthos could control the tectonic movements in the area. Source 4 represents a normal fault, which is located in the coastal area of Cephalonia. The fault dimensions were computed according to empirical relationships (Papazachos and Papazachou, 1997). FAUST database (after Stucchi et al., 2001) suggests this source as the causative fault of the 1636 earthquake (M=7.1).

Figure 10 shows snapshots of the tsunami propagation corresponding to the activation of the Cephalonia North fault (source 4). The corresponding mareograms at the locations depicted in Fig. 4 are presented in Fig. 11.

Judging from the above figures, we can see that the results obtained are similar to those corresponding to the activation of source 3. The greatest wave amplitude is 0.5 m and is reached at Laganas Bay at 20 and 28 min after the source activation.

6 Conclusions

The main objective of this work was the study of possible tsunamis in the Ionian Sea. This was carried out by using a forward modelling approach, that is, by running numerical tsunami simulations, relative to different genetic mechanisms that seem plausible on the basis of the present knowledge of the seismotectonics of the region.

Special emphasis was given to eleven tourist resorts, where we computed the corresponding activation of each of the four sources, mareograms. From all the examined sources, only source 1 possesses a serious threat, causing a wave propagation amplitude that reaches up to 4 m at two locations. Since these locations are heavily populated during the summer season, the need for special tsunami risk mitigation measures is obvious.

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