Historical co-seismic uplift rates in the eastern Hellenic subduction zone: the case of Rhodes island

2 authors:

Ioanna Triantafyllou
National and Kapodistrian University of Athens

Gerassimos A. Papadopoulos
International Society for the Prevention and Mitigation of Natural Hazards

This is a uncorrected version of the paper published in Zeitschrift für Geomorphologie, Vol. 63/2–3 (2022), 201–217, Article Published online August 18, 2021; published in print February 2022

Historical co-seismic uplift rates in the eastern Hellenic subduction zone: the case of Rhodes island

Ioanna Triantafyllou [1,2]* and Gerassimos A. Papadopoulos [2]

[1] Department of Dynamic, Tectonic and Applied Geology, Faculty of Geology and Geoenvironment, National & Kapodistrian University of Athens, Athens 15784, Greece, ioannatriantafyllou@yahoo.gr

[2] International Society for the Prevention and Mitigation of Natural Hazards, Athens, Greece

Corresponding author: ioannatriantafyllou@yahoo.gr

with 7 figures and 2 tables

Abstract. In the eastern Hellenic Subduction Zone important coastal uplift occurred along the east coast of Rhodes Isl. (Greece) during the late Holocene, with uplift gradient increasing from south to north. Various authors estimated geologically that the mean uplift rate has been at ~1 mm a^{-1} in the last ~6 ka and at the range 1.4-3.9 mm a^{-1} in the last 33 ka. To estimate the rate of net, co-seismic costal uplift during the last ~2,3 Ka we analyzed the historical earthquake and tsunami activity based on little-known documentary sources of the 2nd and 19th centuries, on recalculation of magnitudes for some of the earthquakes and eventually on the critical evaluation of the various seismic catalogues available. Published data regarding elevated shorelines, ^{14}C dating and tsunami sediment deposits have been taken into account too. We performed field observations to verify co-seismic uplift in elevated shorelines and in the ancient harbor of Rhodes.

It was found that since the 3rd century BC co-seismic coastal uplift at NE Rhodes has been produced by four non-tsunamigenic earthquakes in ~227 BC, AD~142, 1481 and 1856. The first three shallow earthquakes had their sources very likely close to NE Rhodes, while the 1856 event very likely was an intermediate-depth earthquake. Maximum uplift of ~3.8 m has been attributed alternatively to the 227 BC or to the AD 142 events but the uplift caused by the rest events has been of smaller amplitude. Due to the uncertainties involved in the timing and amplitude of uplift the estimated cumulative net uplift ranges from 5.75 m to 6.95 m, which returns uplift rate of 2.5-3.0 mm a^{-1}. This rate exceeds the late Holocene rate but is consistent with the long-term geological uplift rate. About half of the total seismic moment released by the four episodes was released by one major earthquake (M~7.7) occurring either in 227 BC or in AD 142. Such a major event produced ~60% of the total net uplift. The rest amount of uplift, however, was contributed by smaller but still large (M≥7.0) earthquakes. Coastal uplift contributed by either sediment underplating or aseismic slip is not evident at least during the late Holocene.

Keywords: Co-seismic movements, eastern Hellenic arc, Rhodes Isl., late Holocene co-seismic uplift, historical co-seismic uplift

1 *Introduction*

Coastal uplift and subsidence are geomorphological indicators of tectonic processes and deformation in the upper crust. In convergent plate margins such indicators may also offer insights for better understanding subduction processes and the generation of large earthquakes (e.g. PIRAZZOLI & GRANT 1987, NELSON & MANLEY 1992, SAVAGE et al. 2014, KITAMURA et al. 2019). The Hellenic arc and trench system (Fig. 1) is a plate margin, which is the most active tectonic structure in the Mediterranean Sea region. This is due to the subduction of the Nubian lithosphere beneath the Aegean Sea area (e.g. BOCCHINI et al. 2018 and references therein). From several approaches, including the study of uplifted marine terraces, 2-D viscoelastic modeling and sedimentary correlations, a large variety of uplift estimates ranging from 0.2 to 7.7 mm a^{-1} over timescales since the late Quaternary (~600 ka) up to the present was found in the western and south coasts of Crete Isl., at the western segment of the Hellenic Subduction Zone (HSZ) (e.g. MOUSLOPOULOU et al. 2015, ROBERTSON et al. 2019). The general uplift of the coastal zone seems to be controlled by offshore normal faults (e.g. SKOURTSOS et al. 2017, GALLEN et al. 2014).

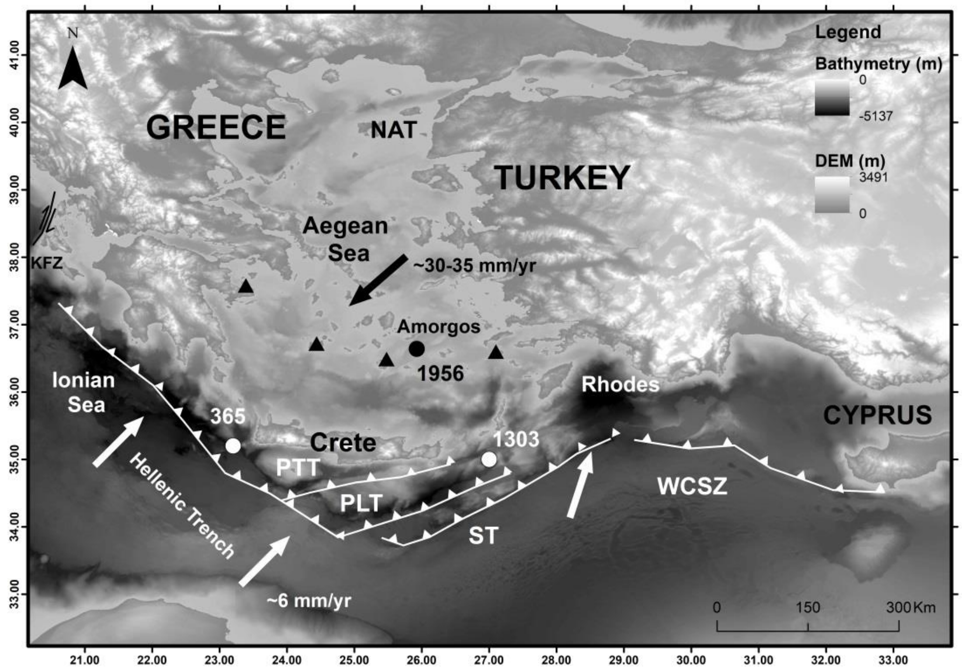

Fig. 1. Map view of the Hellenic subduction zone. Lines with white triangles indicate the deformation front. The Ptolemy (PPT), Pliny (PLT) and Strabo (ST) trenches are also illustrated. Stars show epicentral locations of some of the largest tsunamigenic earthquakes that have occurred in the region: AD 365, 1303, 1956. Arrows and nearby figures show the direction and velocity of motion of the Nubian Plate from about SW to NE (white) and of the Aegean region from NE to SW (black) with respect to the stable Eurasian Plate. Solid triangles indicate the position of the main volcanic centers in the South Aegean volcanic arc. KFZ=Kefalonia Fault Zone; NAT=North Aegean Trough.

Field observations and numerical models indicated that late Holocene uplift along the Hellenic plate margin is primarily achieved by great earthquakes. A well-documented case is the AD 365 tsunamigenic earthquake (M~8.3) that ruptured the western HSZ (e.g. GUIDOBONI et al. 1994, AMBRASEYS 2009, PAPADOPOULOS 2011) and caused a ~9-10 m uplift in SW Crete (PIRAZZOLI et al. 1982, 1996; SHAW et al. 2008). This event likely occurred on a major reverse fault in the upper plate with little contribution from plate-interface slip (SHAW et al. 2008, GANAS &PARSON 2009, MOUSLOPOULOU et al. 2015). To the south of Crete a modern earthquake of this type but of smaller size has been the one of 1 July 2009 measuring M_w=6.5 (BOCCHINI et al. 2019). A similar event of M_w=6.6 ruptured the same area on 2 May 2020. These two events caused also local tsunami waves.

In the eastern HSZ, evidence for uplift in Rhodes Isl. comes from marine sediment distribution and flights of Late Pliocene and Quaternary marine terraces, up to ~250 m high (MUTTI et al. 1970, MEULENKAMP et al. 1972, HANKEN et al. 1996) and from palaeomagnetic stratigraphy indicating 400-600 m uplift in the last ~3 Ma (LØVLIE et al. 1989). Quaternary tectonics on land is associated mainly with uplift motion and normal faulting (MUTTI et al. 1970; GAUTHIER 1979). Late Holocene uplift in east Rhodes is evident from elevated notches radiocarbon dated in the last ~6 ka and taking maximum amplitude of ~3.8 m at the NE tip of the island (PIRAZZOLI 1988, PIRAZZOLI et al. 1989, HOWELL et al. 2015; see also the neotectonic map by LEKKAS et al. 1993). The average coastal uplift increases from 0 at the SE tip of the island to ~1 mm a^{-1} at the NE side (FLEMMING 1978, PIRAZZOLI et al. 1989). Archaeological evidence for late Holocene uplift and subsidence comes from a ~2400-yr-old harbor in Rhodes (BLACKMAN et al. 1996, STIROS & BLACKMAN 2014).

It has been supported that the east coast of Rhodes is characterized by small tectonic fragments with maximum length of ~12 km each one (PIRAZZOLI et al. 1989). KONTOGIANNI et al. (2002) considered that the tectonics is dominated by a main compressional fault zone offshore, sub-parallel to the east coast of Rhodes, with normal faults inland. This result, however, does not imply that the late Holocene uplift should be attributed to only one single major earthquake. On the other hand, HOWELL et al. (2015) supported that the late Quaternary uplift of Rhodes could plausibly be caused by one of the three processes: (1) infrequent large earthquakes, (2) frequent small earthquakes, and (3) continuous gradual uplift related to sedimentary underplating. HOWELL et al. (2015) were based on AMS ^{14}C dates from uplifted paleoshorelines and supported that uplift in late Holocene is most consistent with a single reverse-faulting earthquake of M≥7.7 occurring between almost certainly after 2000 BC, and possibly after 300 BC. However, in the discussion about the timing and size of late Holocene uplift in east Rhodes a thorough examination of the historical seismicity is nearly lacking. Only limited consideration has been given to historical documentation of co-seismic tectonic movements (e.g. KONTOGIANNI et al. 2002, STIROS & BLACKMAN 2014).

Three main questions still remain unresolved regarding the late Holocene uplift in NE Rhodes. (i) Has the uplift been caused by a single major earthquake? (ii) Is it possible to estimate reliably the historical uplift rate? (iii) Is the last consistent with rates estimated in late Holocene? To reply these key issues we examined the historical seismicity record of the area and performed coastal field trips to verify the late Holocene uplift elevation at some key coastal spots. Yet, we inspected an archaeological excavation providing evidence of tectonic movements in the ancient harbor of Rhodes. Eventually, taking into account also published results the historical co-seismic uplift rate was estimated and compared with the geological uplift rate in the late Holocene.

2 *Geodynamic and seismotectonic setting*

Rhodes occupies the easternmost side of the HSZ (Fig.1). In late Miocene-Pleistocene, the outward migration of the absolute position of the convergent Hellenic plate boundary produced simultaneous increased curvature of the boundary, changing obliquity of plate convergence vectors and boundary-parallel stretching of the fore-arc region causing WNW-ESE extension (283^0) and 070^0 left-lateral shear (TEN VEEN & KLEINSPEHN 2002). Magneto-biostratigraphy and $^{40}Ar/^{39}Ar$ dating of a Plio-Pleistocene volcaniclastic layer in the central-east coast showed (1) 500-600 m drowning during the latest Pliocene, (2) at least 520 m of uplift at ~1.4-1.3 Ma, (3) counterclockwise motion of Rhodes, younger than 1.2-1.1 Ma (CORNÉE et al. 2006). Tilting towards the Aegean (KONTOGIANNI et al. 2002) or rotations about a vertical axes (SHAW AND JACKSON 2010) possibly are involved in the kinematics of the area.

Examination of Middle-Upper Pleistocene deposits of Rhodes showed that the eastern coast has experienced two major uplift episodes, between 800 and 400 ka and since 33 ka, punctuated by subsidence phases (CORNÉE et al. 2018). Uplift and subsidence rates were rather high at 0.32–0.5 to c. 2–2.4(?) mm a^{-1}, i.e. in the range classically observed in other fore-arc settings. From 33 ka the uplift rate ranged between 1.4 and 3.9 mm a^{-1}, coeval with a 17±9° anticlockwise rotation. According to geodetic measurements the Aegean region/Nubian plate convergence rate (~35 mm a^{-1}) exceeds the Africa/Eurasia convergence rate (~5–10 mm a^{-1}). This is attributed to the rapid SW-motion of the Aegean region, with velocity increasing towards the trench with respect to Eurasia (McCLUSKY et al. 2000, REILINGER et al. 2006, 2010; Fig. 1).

Seismic reflection indicated the existence of reverse faults between Rhodes and Rhodes Basin (Fig. 2) dipping to NW (WOODSIDE et al. 2000, HALL et al. 2009). Relocated earthquake hypocentres showed a NW dipping subduction slab down to ~180 km (BOCCHINI et al. 2018). The Rhodes block moves to southeast at rates up to ~10 mm a^{-1} with respect to the Anatolia (REILINGER et al. 2006). Large earthquakes in the area of Rhodes have been documented from the antiquity up to the present (Fig. 2; see references in next section). Earthquake focal mechanisms indicate along-strike (NE-SW) compression and active shortening acting on the Nubian slab (TAYMAZ et al. 1991, BENETATOS et al. 2004, SHAW & JACKSON 2010). From good quality relocated seismicity (BOCCHINI et al., 2018) a tearing of the subducting slab was found between Rhodes and SW Turkey, with the Eastern HSZ slab subducting towards NW underneath Rhodes and the Western Cyprus Subduction Zone (WCSZ) with subduction direction towards NE beneath SW Turkey.

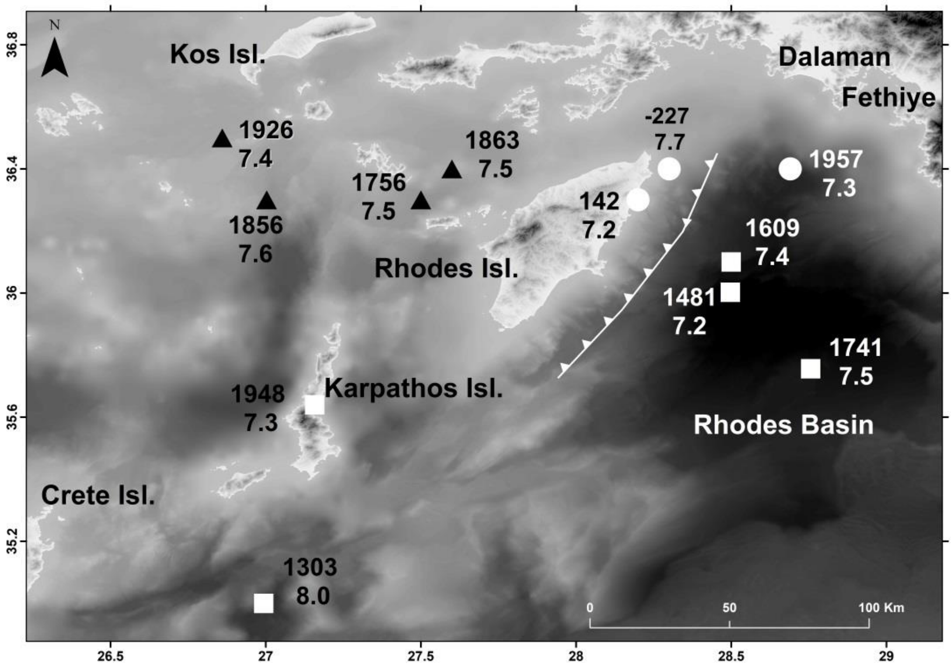

Fig. 2. Earthquake epicenters, years (- means BC) and magnitudes of historical earthquakes in the area of Rhodes Isl. Key: circle =shallow non-tsunamigenic earthquake; square= shallow tsunamigenic earthquake; triangle=intermediate-depth earthquake. Line with white triangles indicate reverse faults lying between Rhodes and Rhodes Basin about perpendicular of the subduction direction. For references see the text.

3 *Method and data*

The sea-level observed at a coastal site at a given point of time depends on co-seismic movements, on eustatic sea-level changes and on tidal variation. Cycles of tectonic uplift-subsidence have been described along the NE coast of Rhodes (e.g. PIRAZZOLI et al. 1989, KONTOGIANNI et al., 2002, STIROS & BLACKMAN 2014). At all evidence the eustatic sea level in the eastern Mediterranean in the last ~6 ka has been nearly stable not being higher than the present level (e.g. LAMBECK 1995, SIDALL et al. 2003, SIVAN et al. 2004). The tidal range in the area is usually less than 0.2 m (PIRAZZOLI et al. 1989). Therefore, we estimated the net magnitude of coastal co-seismic uplift rate in NE Rhodes during historical times and compared it with published results on the late Holocene uplift rate in the same area.

Correlating historical, geological and radiometric data is a productive approach to study correlations between late Holocene seismicity and tectonic movements (e.g. KITAMURA et al. 2019). Our investigation has been based on two sets of observational data indicating co-seismic coastal displacements in the eastern side of Rhodes. The first set of data concerns the historical seismicity record in Rhodes, which is extended up to the 3rd century BC. This data set includes two documentary sources which have not been used in the seismological tradition so far. The first is a little-known document of the 2nd century AD about the large earthquake of ~ AD 142. The second document is an unpublished text of the 19th century AD referring to the destructive earthquakes of AD 1851 and 1856. In addition, parametric earthquake catalogues compiled and published by several authors have been taken into account for the entire time period examined. Parameters of the earthquakes examined are summarized in Table 1. These parameters have been adopted from several seismicity catalogues but magnitudes for some earthquake events have been calculated by us as explained below. Tsunamis are indicators of co-seismic sea-floor displacements and for this reason such events have been considered in our analysis too.

The second data set comprises published geological, radiometric and archaeological observations and results indicating co-seismic tectonic movements in the last ~6 ka. In addition, we performed field observations, measured amplitudes of elevated paleoshorelines in several coastal spots of eastern Rhodes and compared them with published results.

3.1 Historical earthquakes and tsunamis

3.1.1 Historical catalogues

Documentary sources regarding historical earthquakes and tsunamis that occurred in the study area have been compiled, reproduced and utilized by several authors who organized local or regional, descriptive and/or parametric earthquake and tsunami catalogues, including GALANOPOULOS (1960), AMBRASEYS (1962, 2009), PAPADOPOULOS & CHALKIS (1984), GUIDOBONI et al. (1994), PAPAZACHOS & PAPAZACHOU (1997, 2003), PAPADOPOULOS et al. (2007, 2014), GUIDOBONI & COMASTRI (2005), PAPADOPOULOS (2011) and MARAMAI et al. (2014). The earthquake catalogue by PAPAZACHOS et al. (2010) is a copy of the parametric part of the PAPAZACHOS & PAPAZACHOU (2003) descriptive and parametric catalogue. Of interest is also the earthquake catalogue produced by the EC SHARE project (STUCCHI et al. 2013) as well as the EMEC catalogue (GRÜNTHAL & WAHLSTRÖM, 2012, GRÜNTHAL et al., 2013), both starting from 1000 AD. The EMEC catalogue relies extensively on the catalogue by PAPAZACHOS & PAPAZACHOU (2003). The CFTI5Med catalogue (Guidoboni et al., 2018, 2019) is the last version of previous CFTI5Med catalogue versions but for Greece covers only the time period from the antiquity up to 1500 AD. The book by PAPADOPOULOS (2016) is an exhaustive review of the earthquakes and tsunamis that occurred in the area of Rhodes from the antiquity up to 2016.

3.1.2 Historical documents

Apart from the documentation of historical earthquakes and tsunamis that can be found in the papers, books and databases quoted in the previous subsection, of particular interest are the next two documents: (1) *Rhodian Oration* and (2) *Geschichte der Insel Rhodos*. The content of these two sources is analyzed below but their seismotectonic implications are explained in section 3.3. The *Rhodian Oration* is a little known document written in Greek and ascribed to the formidable sophist Aelius Aristides (AD 117-180). It is of great value for better understanding the earthquake that isolated Rhodes in ~AD 142. It is at once a commemoration of the ruined city, a memorial of the catastrophe, and an exhortation to the survivors (FRANCO 2008). The earthquake destructive effects are vividly described in a section of the speech (§17–33). Aristides collected information from others and likely visited Rhodes by himself shortly after the event. GUIDOBONI et al. (1994) ignored that document but AMBRASEYS (2009) published an English translation of only some passages of it.

An equally important document is the unpublished manuscript *Geschichte der Insel Rhodos* (*A History of Rhodes Island*) written in German, from 1854 to about 1863, by the Swedish medical doctor Johannes Hedenborg (1786-1865). As a permanent resident at Rhodes, J. Hedenborg witnessed several earthquakes including the AD 1851 and 1856 ones, which are of interest to our analysis. We were able to read the manuscript twice on 17 June 2012 and on 30 January 2015.

Table 1 lists parameters of the significant earthquakes that are known to have happened in the area of Rhodes from the antiquity up to the present. In our analysis the term "significant earthquake" refers to an event which is of estimated magnitude 7 or larger in at least one earthquake catalogue. The only exception is the earthquake of 18 December 1481 of estimated magnitude of ~6.5. This earthquake is included in Table 1 since it has likely been associated with a coastal uplift. This case is explained later. It is noteworthy, however, that in several cases the magnitudes estimated in the various catalogues for an individual earthquake are quite divergent. Therefore, magnitude M in Table 1 is the preferred magnitude, while M* indicates other magnitude estimates. Magnitude M of the 227 BC, AD 142, AD 1481 and AD 1609 earthquakes have been calculated by us from an empirical relation between M and maximum displacement for reverse faults (WELLS & COPPERSMITH 1994; see text). The earthquake of 28 February 1851, which caused remarkable damage in Rhodes (e.g., HEDENBORG unpublished) but mainly to SW Turkey, has eventually not been included in the list of Table 1 for the reason that in the earthquake catalogues available (PAPAZACHOS &

PAPAZACHOU 2003, PAPADOPOULOS 2011, SCHEEC 2013) its estimated magnitude is less than 7.

Table 1. Significant historical earthquakes in the area of Rhodes. Key: y=year, m=month, d=day, ϕ^o_N and λ^o_E=geographical coordinates, M=magnitude proxy M_w, h=focal depth (km), n=normal, ni=interplate, i=intermediate-depth, co-ph=co-seismic phenomena, u=coastal uplift (in m), symbol / means "or", t=tsunami, M*=other magnitude, R=NE Rhodes Isl.; He=Heraklion, Crete; D=Dalaman, SW Turkey; K=Karpathos. Parameters are estimated in this paper unless otherwise indicated in parenthesis: (1) PAPAZACHOS & PAPAZACHOU 2003, (2) AMBRASEYS 2009, (3) ISC-GEM 2020, (4) PAPADOPOULOS 2011, (5) AMBRASEYS & ADAMS, 1998, (6) SHEEC catalogue, (7) MAKROPOULOS et al. 2012, (8) AMBRASEYS 2001, (9) HOWELL et al. (2015), (10) CFTI5Med catalogue. Magnitude, M, estimated by us has been based on empirical relations between M and maximum fault displacement, U (WELLS & COPPERSMITH 1994).

Date y m d	φ_N^o	λ_E^o	M	h	co-ph	M*
227BC	36.4	28.3	7.7/7.2	n	uR 3.8/1.5	7.5(1), ≥7.7(9), 6.6(10)
198BC	36.17 (10)	28.00 (10)	6.6(10)	n		7.2 (1)
AD142	36.3	28.2	7.2/7.7	n	uR 1.5/3.8	7.5 (1)
1303 08 08	35.0 (4)	27.0 (4)	8.0±0.3(4)	ni (4)	t(He,D)	8(6), 8.26±0.3(6), 8.8(10)
1366 04 30	36.43 (6)	28.23 (6)	6.75±0.3(6)			7.2 (1)
1481 05 03	36.0	28.5	≥7.2	n	t(R,D)	7.2 (1), 6-7 (2)
1481 12 18	36.2	28.3	6.5	n	uR1.1	6.42±0.3(6), 5.8 (10)
1513 03 28	36.1 (6)	28.2 (6)	6.21±0.5(6)			7.2 (1)
1609 04	36.1	28.5	~7.4	n	tR	7.2±0.5 (1,6), 7-8 (2)
1741 01 31	35.8	28.8	7-8(2)	n	t(R,D)	7.4 (1), 7.59±0.3 (6)
1756 02 13	36.3 (1)	27.5 (1)	7.5±0.5(1)	i (1)		
1856 10 12	36.3	27.0	7.6±0.3(4)	60(1)	uR 0.55	7.7 (1,6), 7-8 (2)
1863 04 22	36.4 (1)	27.6 (1)	7.5(1)	i		7.5±0.3 (6)
1926 06 26	36.50 (2)	26.86 (2)	7.4±0.3(5)	115(5)		7.6(1),7(7), 6.99±0.2(3)
1948 02 09	35.64 (3)	27.158 (3)	7.3±0.2(3)	15(3)	tK	7.1 (1)
1957 04 24	36.49 (3)	28.82 (3)	7.1±0.2(3)	35(3)		6.8 (1), 6.6 (8), 6.8 (7)
1957 04 25	36.40 (3)	28.69 (3)	7.3±0.2(3)	35(3)		7.2 (1), 7.1 (7), 6.79 (8)

3.2 Geological and archaeological observations

Data on the geological uplift rate in Rhodes during the late Holocene can be found in various publications (FLEMMING 1978, PIRAZZOLI et al. 1989, BLACKMAN et al. 1996, KONTOGIANNI et al. 2002, STIROS & BLACKMAN 2014, HOWELL et al. 2015). During four field trips performed in Rhodes from June 2012 to October 2016, we inspected several sites with uplifted shorelines along the east coast and estimated the magnitude of uplift. In addition, we evaluated the magnitude of uplift and the radiocarbon dates published by other authors and correlated them with historical evidence for co-seismic uplift. Table 2 summarizes the main results of our investigation as well as of other authors, while the map in Figure 3 illustrates the main observations listed in this Table.

Table 2. Summary of co-seismic uplift observations in the eastern coastal side of Rhodes Isl. Relative and net uplift are indicated by r and n, respectively. Radiocarbon date is either BP or calibrated (c). Symbol – in historical date means BC. Key for references: T (present study), P (PIRAZZOLI et al., 1989), H (HOWELL, 2015), S (STIROS & BLACKMAN, 2014).

Site	Lat(°N)	Long(°E)	Uplift elevation a.m.s.l. (m)	Radiocarbon Date	Historical date
Rhodes harbour	36.4494	28.2253	~0.55 (T)		1856 (T)
Rhodes ancient harbour	36.4460	28.2262	~4.0 (S) 3.8 or 1.5 (T)		~200 (S) -227 or 142 (T)
Kalitea (Kallithea)	36° 23′	28° 17′	2.35 (Pr) 3.8 (Pn)	-2280(±110)(P) -178-(+459) (Sc)	-222 (P)
Kalitea (Kallithea)	36° 23′	28° 17′	1.17 (P) 0.74 (P)	855±70 (P) 1336-1682 (Sc) 895±70(P) 1342-1721 (Sc)	1481 or 1609 (T)
Ladiko	39° 19′ 36.32 36.3196	28° 15′ 28.21 28.2112	3.8 (P) 3.8 (H) 3.75 (T)		-227 or 142 (T)
Traganou	36.31 36.3104	28.19 28.1949	3.8 (H) 3.8 (T)		-227 or 142 (T)
Tsampika (Zambica)	36°20′ 36.22 36.2323	28°15′ E coast 28.1580	3.2 (P) 3.0 (Hn) 3.0 (T)	>2000BC, likely>300 BC(H)	-222 (P) -227 or 142 (T)
Charaki	36° 08′	28° 10′	0.3 (P)	1840±60 (P)	142? (T)
Prasonisi SW of Lindos	35.8879 36.09	27.7801 28.09	0 (P, T) ~0 (H)		

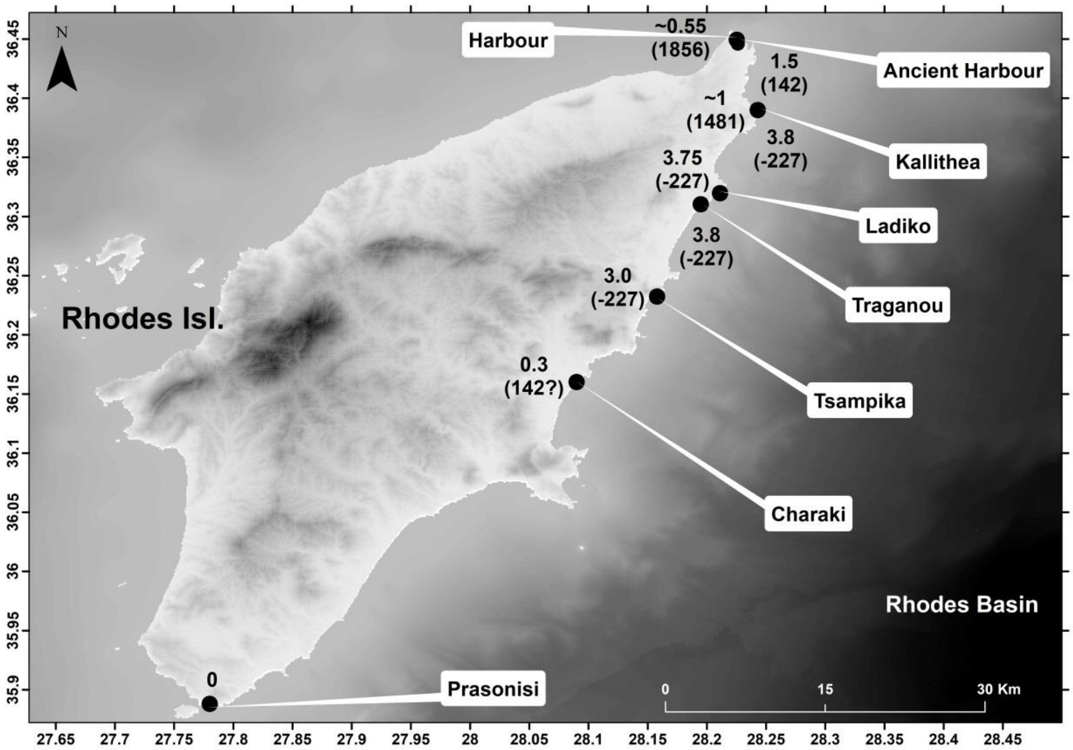

Figure 3. Years of historical earthquakes (- means BC) and associated co-seismic uplift (in m) in east Rhodes coast. Data sources explained in Table 2.

Based on our field observations we were able to verify that moving from south to north uplift of ~0 m is evident in Prasonisi (Fig. 3, 4A). On the contrary, the top raised shoreline in the St. Paul bay of Lindos is ~2.4 m a.m.s.l. (Fig. 4B), in Tsampika is ~3 m (Fig. 3, 5) and in Traganou and Ladikou, NE side, is ~3.8 m a.m.s.l. (Fig. 3, 6A,B). These observations are consistent with those published by others (PIRAZZOLI et al. 1989, HOWELL et al. 2015). Archaeological evidence for co-seismic tectonic movements comes from an excavation at the ancient harbor of Rhodes (BLACKMAN et al. 1996, and STIROS & BLACKMAN (2014). With the permission of the local Archaeological Survey we visited and inspected this excavation on January 30th, 2015.

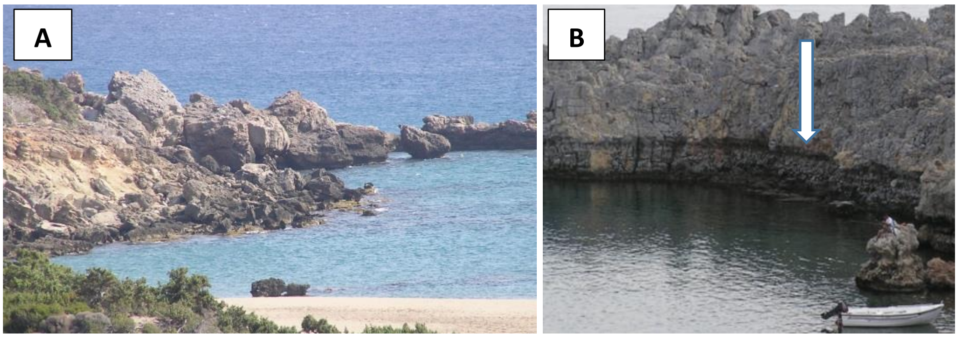

Fig. 4. (A) No evidence of coastal uplift has been found in Prasonisi, southern tip of Rhodes Isl.; (B) Uplift of 2.4 m at St. Paul bay, Lindos, central-east coast (photo courtesy by I. Triantafyllou).

Fig. 5. Coastal uplift of ~3 m in Tsampika, NE Rhodes coast (photo courtesy by I. Triantafyllou).

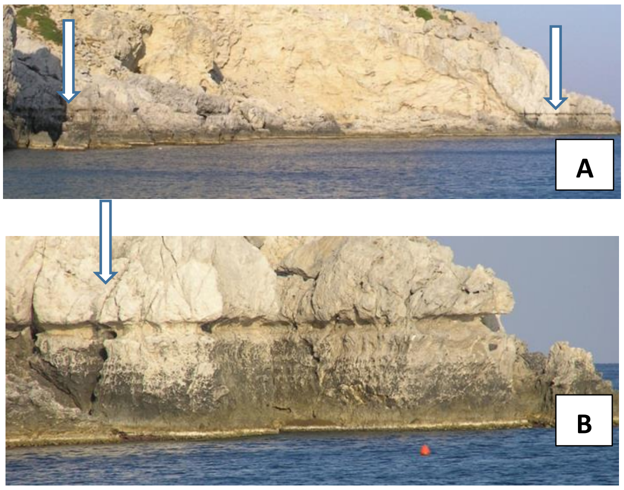

Fig. 6. (A) Coastal uplift of ~3.8 m in Traganou, NE coast of Rhodes Isl.; (B) focus at the paleoshoreline notch at the right-hand side of (A) (photo courtesy by I. Triantafyllou). Uplift of the same amplitude is evident in the nearby Ladikou beach.

Tsunami sediment deposits found in a trench opened in a marshy spot at ~240 m inland at Dalaman, SW Turkey (Fig. 2), were dated in the historical period (PAPADOPOULOS et al. 2012). Three sea sand layers were identified at elevations of +0.30, +0.55 and +0.90 cm. AMS ^{14}C dating of a wood sample from layer II indicated deposition in AD 1473±46, which matches the 1481 tsunamigenic earthquake. From an estimated average alluvium deposition rate of ~0.13 cm a^{-1} layers I and III were dated at about 1322 and 1724, which may represent the large 1303 and 1741 seismic tsunamis, respectively. However, the 1609 tsunami is missing from the stratigraphy in that trench. ALPAR et al. (2012) performed re-sampling in the Dalaman trench and analyzed biomarkers that indicated only the 1481 event as a tsunami deposit since no biomarkers were found in the other two layers. In the nearby Ölüdeniz Lagoon sedimentary geochemical evidence were found for the 1303, 1609 and 1741 seismic tsunamis (AVȘAR 2019).

3.3 *Historical earthquakes and co-seismic uplift*

Historical, geological and radiometric data regarding the significant earthquakes listed in Table 1 are presented and evaluated in the next lines. The various documentary sources available are not presented in details here since they have been published and reviewed by many authors quoted in the previous sections.

The ~227 BC earthquake

Classic authors reported on an earthquake occurring ~227 BC, very likely between 226 and 229 BC. The earthquake isolated Rhodes causing collapse of the city walls and shipyards as well as of the Colossus of Rhodes, one of the seven wonders of antiquity. An inscription dated in the 3rd century BC or later says that the earthquake killed people. The damage was possibly extended in SW Asia Minor (modern Turkey) provinces. No tsunami was reported. In Kallithea, a few kilometers to the south of Rhodes city, the conventional radiometric age of a calcareous algal sampled at an emergent shoreline by 2.35 m was found 2280±110 B.P. (±1σ) by PIRAZZOLI et al. (1989). These authors, considering coastal subsidence and taking as reference level the nearby coastal quarries of very likely Roman era (FLEMMING 1978), which are now slightly submerged, concluded that a rapid 3.8 m fall in the relative sea level, probably corresponding to a sudden uplift movement accompanying the earthquake of 227 BC, happened slightly after the deposition of the sample. For the same sample, however, STIROS & BLACKMAN (2014) found calibrated age 178 BC-AD 459 (±2σ). This result matches the AD 142 earthquake event but leaves out the 227 BC one.

Further to the southwest, HOWELL et al. (2015) used radiometric date estimated from lithophaga sampled at coastal site 0.7 m a.m.s.l. and identified a tectonic event occurring in 2009-1532 BC (±2σ). Incorporation of a 400-yr offset between the radiocarbon age and the death of the organism, as suggested by SHAW et al. (2008) and EVELPIDOU et al. (2012) for late Holocene uplifts in other coastal sites, brings that age range to ~1600-1100 BC (HOWELL et al. 2015), which does not match the 227 BC earthquake but likely represents an earlier tectonic event. A sample of arcoid collected nearby at 1.5 m a.m.s.l. was radiocarbon dated at 316 BC-AD 155 (±2σ). HOWELL et al. (2015) suggested that no offset correction was needed as the sample found to be formed entirely by aragonite, thus its age likely representing the death of the organism. Based on radiometric results, those authors suggested the earliest possible date for the earthquake was 2000 BC, possibly after 300 BC.

Archaeological evidence for tectonic movements comes from slipways in an excavation site at the ancient harbor of Rhodes (BLACKMAN et al. 1996). An older phase of the slipway, corresponding to +2.05-3.10 m a.m.s.l at present was repaved by a later slipway at +2.50-4.05 m. This was attributed to the destruction of the first slipway level by the 227 BC earthquake, thus causing need for reconstruction at a level of ~1 m higher possibly during the 2nd century BC. KONTOGIANNI et al. (2002) and STIROS & BLACKMAN (2014) suggested that the only possible explanation for the reconstruction was to counteract a transient seismic land subsidence, the tectonic origin of which could be attributed to the 227 BC earthquake. From our inspection of the excavation, however, it comes out that neither the top nor the bottom of the slipway can be identified, which implies that the amplitude of the co-seismic ground displacement is not well constrained.

We suggest that the large 227 BC earthquake had its source close to Rhodes (Fig. 2), very likely in the shallow water domain, which may explain the non-tsunamigenic nature of the event. Coastal uplift in NE Rhodes associated with this earthquake was of 3.8 m in amplitude. An alternative is that uplift of 3.8 m was associated with the AD 142 event, then uplift amplitude of only 1.5 m occurred in 227 BC.

The ~142 AD earthquake

According to ancient authors and inscriptions, Rhodes was isolated again by a large earthquake in ~AD 142, which likely affected the island of Kos and SW Asia Minor (Fig. 2). The *Rhodian Oration* constitutes a key document for understanding the earthquake. AMBRASEYS (1962) considered that *Rhodian Oration* provides evidence for a strong tsunami accompanying the earthquake and supported that Rhodes was ruined by the tsunami (AMBRASEYS 2009). However, such a conclusion is unfounded. We examined the original Greek text, translations of it in modern Greek (e.g. PAPAIOANNOU 1995) and an English version of the relevant passages as published by AMBRASEYS (2009) himself, which read as follows: …*The sea drew back, and all the interior of the harbours was laid bare, and the houses were thrown upwards…In the time that it took for a man to raise anchor to sail off, when he turned around, he could no longer see the city, but everything was jumbled together, the harbours on dry land, the city in the dust, empty streets…Everything happened at once: the earthquake from the sea, the cloud, the noise, the cries, and the crash of the ruins, the heaving of the earth…*

The seismotectonic implications are of great importance. First, the *Rhodian Oration* does not imply at all that a tsunami was generated since *the sea drew back* but without return and inundation onshore as it happened with the AD 365 large tsunami in Alexandria (see

Ammianus Marcellinus/AD 322-395; L.XXVI, 10.15-19, in GUIDOBONI et al. 1994). Second, that the interior of the harbours was laid bare but without tsunami generation provides evidence for permanent co-seismic uplift. This is further supported by that *the houses were thrown upwards* and *the harbours [became] dry land* because of *the heaving of the earth*. Besides, the statement *the earthquake [came] from the sea* implies no tsunami generation but an offshore seismic source. Should the city had ruined because of a tsunami it very likely would not escape historical record.

In Charaki, ~30 km to the SW from Rhodes city, the sea level dropped in 1840±60 yr (±1σ) B.P., when an algal *N. notarisii*-bearing crust developed up to +0.3 m (PIRAZZOLI et al. 1989). From archaeological observations, indicating tectonic movements from slipways at the ancient harbor of Rhodes around ~230 BC, BLACKMAN et al. (1996) and STIROS & BLACKMAN (2014) suggested that ~300 years later the harbor was to a great part abandoned due to coastal uplift. Perhaps this is the uplift inferred from *Rhodian Oration* and radiocarbon dating. Little doubt remains that a large, non-tsunamigenic earthquake destroyed Rhodes around AD 142. We concluded that independent historical, paleoshoreline and radiometric data indicate that very likely the earthquake was associated with coastal uplift. The uplift amplitude ranges from ~0.3 m about 30 km to the south of Rhodes city, to 1.5 m further to the north and to 3.8 m close to the NE tip of the island, if the last is not associated with the 227 BC earthquake. Otherwise uplift amplitude of only 1.5 m occurred in AD 142.

The 1303 earthquake

The large (M~8) 1303 earthquake (Fig. 2) caused widespread destruction in Crete and Rhodes. AMBRASEYS et al. (1994) suggested that Rhodes was hit by the 1303 tsunami. One of the tsunami sediment deposits found in Dalaman (PAPADOPOULOS et al. 2012) and in the nearby Ölüdeniz Lagoon (AVȘAR 2019), SW Turkey, was attributed to the AD 1303 tsunami. No coastal uplift has been documented probably due to that the 1303 causative fault outcrops in either the Pliny or Strabo Trench (Fig. 1) (e.g. HOWELL et al. 2015). The two thrust faulting earthquakes of 1 July 2001 (M_w=6.4) and of 2 May 2020 (M_w=6.6) had their sources in the Pliny/Strabo Trench system and produced small tsunamis (BOCCHINI et al., 2020, PAPADOPOULOS et al., 2020).

The 1481 seismic sequence

A persisting seismic sequence was felt in Rhodes during 1481. The first strong shock of 18 March 1481 caused considerable damage in Cyprus but not in Rhodes (GUIDOBONI et al. 1994). The seismic source likely was associated with the WCSZ away from Rhodes. The next strong earthquake of 3 May 1481, although did not damage Rhodes, yet it triggered a ~3 m-high-tsunami that reached up to the middle of the city and destroyed a boat that was anchored by. We propose a source at considerable distance from Rhodes and very likely at deep sea water, likely in the area of Rhodes Basin (Fig. 2). AMS ^{14}C calibrated dating of the tsunami sediment layer II found at Dalaman indicated deposition in AD 1473±46 (±1σ), which fits fairly well the 3 May 1481 event (PAPADOPOULOS et al. 2012). Sand dykes directed upwards from layer I to layer II indicated that the 1481 earthquake triggered liquefaction of sand layer I, the one that very likely left behind by the 1303 tsunami. We applied empirical M/R relationships predicting the maximum distance, R, at which soil liquefaction may occur as a function of magnitude, M (PAPADOPOULOS & LEFKOPOULOS 1993, PAPATHANASIOU et al. 2005) and found that M=7.2 is sufficient to get R~125 km, which certainly exceeds the distance from Rhodes Basin to Dalaman coast. Then, M=7.2 is a lower magnitude threshold for this earthquake. The third strong earthquake of 18 December 1481 hit Rhodes causing the collapse of churches, palaces and castles as well as many human victims. Ground fissures were opened but no tsunami was reported. We suggest that the source of the 18 December 1481 earthquake should be placed close to Rhodes (Fig. 3).

Near Kallithea, a few kilometers to the south of Rhodes city, fossil barnacle samples collected from ca. +1.17 and +0.74 a.m.s.l. have been dated 855±70 (±1σ) yr B.P. and 895±70 (±1σ) yr B.P. (PIRAZZOLI et al. 1989). However, STIROS & BLACKMAN (2014) found calibrated date ranges AD 1336-1682 (±2σ) and AD 1342-1721 (±2σ). It should not rule out that these dates may represent a coastal uplift phase of ~1.0 m related likely to the 18 December 1481 earthquake. Then, from the uplift amplitude we calculated earthquake magnitude equal to ~6.5 (Table 1). The other candidate earthquakes of 3 May 1481 and April 1609 have been tsunamigenic, thus implying sources at rather deep water domain, e.g. in Rhodes Basin away from Rhodes coasts. In either case our estimation of the historical uplift rate does not change.

The 1609 and 1741 earthquakes

A large tsunamigenic earthquake caused destruction at Rhodes city in April 1609:...*half of the town, including the castle was ruined, and [an exaggerated figure of] over 10000 people were reported drowned by a sea-wave...This appears to have been a great earthquake, felt also in various places in Egypt and the Syrian coast but further details are lacking* (AMBRASEYS et al. 1994). If the tsunami generation implies earthquake source in Rhodes Basin (Fig. 2), then large magnitude is required to cause extensive damage in Rhodes.

On 31 January 1741 Rhodes was hit by another large earthquake that caused severe damage in houses and walls of the city, while damaged Cyprus too. A damaging tsunami was reported: *...the sea in Rhodes retreated and then flooded the coast 12 times with great violence, submerging the coast opposite the island and destroying five or six villages located a kilometer inland* (AMBRASEYS et al. 1994). The upper tsunami layer found at Dalaman, SW Turkey, very likely represents geological signature of the 1741 tsunami (PAPADOPOULOS et al. 2012). The seismic source should be associated with active faults either in Rhodes Basin (Fig. 2) or even further in the WCSZ.

The 1756 and 1856 earthquakes

On 13 February 1756 a very strong earthquake, possibly of intermediate depth and with its source in the eastern Hellenic Arc, shook large part of the eastern Mediterranean and caused moderate damage in Rhodes. There is no evidence for tsunami and co-seismic uplift.

A well-known large, intermediate-depth (h~60 km) earthquake occurred in the Cretan Sea (Fig. 2) on 12 October 1856 with perceptibility area and damage zone extending up to Malta, ~1000 km away from the source (see reviews by AMBRASEYS 2009, PAPADOPOULOS 2011). More than 600 people were killed, about 538 only in Crete. No tsunami was reported. According to HEDENBORG (unpublished ms.) and SCHMIDT (1879) the earthquake caused extensive damage in Rhodes where about 60 people killed. HEDENBORG (unpublished manuscript) was an eyewitness of the earthquake in Rhodes and his document contains valuable observations. Of particular interest is the next account: …*The sea drew back with the first seismic motion and after some months had not returned to its normal place…until April 1857 the sea level was yet the same. The difference was 1½ to 2 French feet* [c. 50-60 cm]. *Did the island uplifted as it happens with some earthquakes? Besides, the fact is that Rhodes Isl. uplifted gradually by 2½ French feet* [c. 80 cm] *in the last 60 years, whereas the coast of Anatolia to the opposite site of the island subsided.* This description implies that a co-seismic uplift of ~0.55 m caused permanent drop of the sea level. Probably this uplift episode represents the drop by 30-40 cm of the very recent sea-level A_7 identified close to Rhodes by PIRAZZOLI et al. (1989). Assuming magnitude M=7.6-7.7 (Table 1) the uplift amplitude is relatively small, which, however, is attributed to the deep source.

The 1863 and 1926 earthquakes

On 22 April 1863 a large earthquake (Fig. 2) caused extensive destruction in Rhodes city as well as in many villages of the island; about 224 people killed. Neither tsunami no coastal uplift have been reported and this may imply source at intermediate-depth which, however, is weakened by the intense aftershock activity. Therefore, an interplate rupture has been proposed (PAPADOPOULOS 2011). Another large (M_s=7.4) intermediate-depth earthquake with source placed offshore NW Rhodes (Fig. 2) occurred on 26 June 1926. Extensive damage was caused in Crete. In Rhodes only two people killed although 2000 out of 10000 houses damaged beyond repair. Neither tsunami nor coastal uplift were reported.

The 1948 and 1957 earthquakes

The 9 February 1948 shallow, large (M_w=7.3) earthquake (Fig. 2) caused a local damaging tsunami in Karpathos Isl. (GALANOPOULOS 1960, PAPADOPOULOS et al. 2007). No coastal uplift was reported in association with this earthquake which had its source at considerable distance from Rhodes anyway.

The last large earthquake (M_w=7.3) in the area of Rhodes occurred on 25 April 1957 with epicenter near Fethiye (Makri), SW Turkey (Fig. 2). A very strong (M_w=7.1) foreshock occurred the day before. The main zone of destruction was noted in the area of Fethiye. Neither tsunami nor coastal uplift were reported.

4 Results

Summarizing historical, geological and radiometric results we concluded that co-seismic uplift in NE Rhodes has been caused by the seismic episodes of ~227 BC, ~AD 142, 18 December 1481 and 12 October 1856. Coastal uplifts of 3.8 m and 1.5 m, observed in two nearby sites, fall in radiocarbon dated time windows matching both the 227 BC and the AD 142 earthquakes. Further to the south an episode of +0.3 m correlates also with the AD 142 event. A coastal uplift phase of ~1 m is very likely related with the 18 December 1481 earthquake, while permanent coastal uplift by ~0.55 m was reported after the large earthquake of 1856, which has been considered as an intermediate-depth one. In view of the uncertainties involved in the uplift amplitudes mainly associated with the 227 BC and AD 142 events, the average co-seismic uplift in NE Rhodes Isl. during the last ~2.3 ka is estimated at 2.5-3.0 mm a^{-1}. Empirical relations between magnitude, M, and maximum fault displacement, U, returned M~7.7 (WELLS & COPPERSMITH 1994) or M~8.3 (PAVLIDES & CAPUTO 2004) for U=3.8 m. The seismic moment, M_o (HANKS & KANAMORI 1979; formula 1), released by a major earthquake of M~7.7 is nearly half of the total moment released by the four seismic episodes (Fig. 7):

$$\log_{10} M_o = 1.5\ M + 16.05 \quad (1)$$

One major (M≥7.7) earthquake, either that of 227 BC or alternatively the AD 142 one, supplied about 55-65% of the total historical co-seismic uplift. The rest part of uplift was supplied by smaller but still large magnitude (M >7) earthquakes (Table 1, Fig. 7).

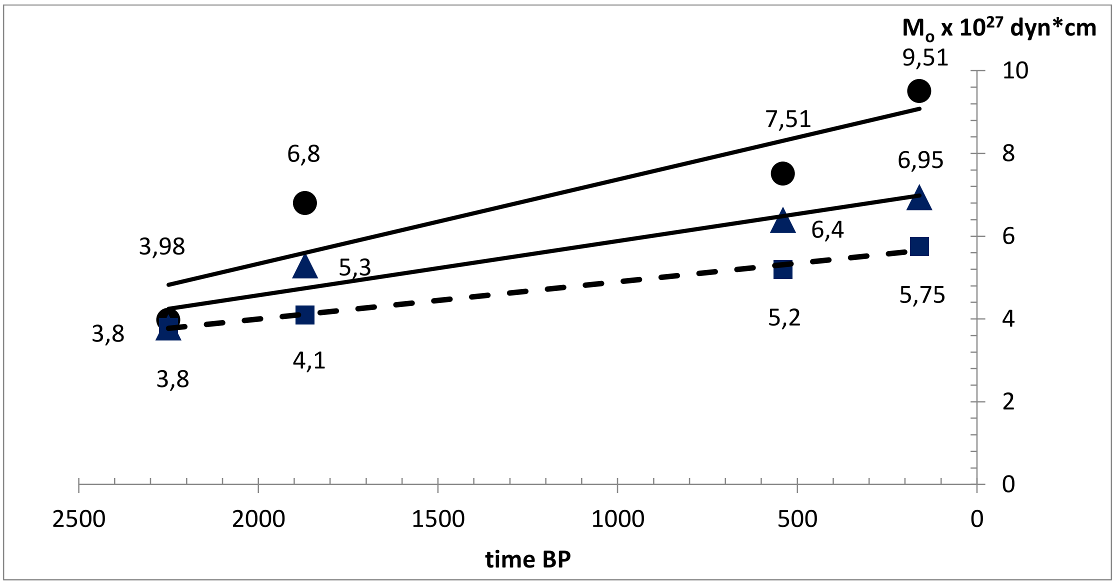

Fig. 7. Cumulative historical coastal uplift (in m, solid triangle; solid square shows conservative estimation) caused in NE Rhodes by the 227 BC, AD 142, 1481 and 1856 earthquakes, and seismic moment (x 10^{27}dyn.cm, solid circle) released by these earthquakes. Time is in years before present time 0. We assumed that the maximum uplift amplitude of 3.8 m is associated with the earthquake of 227 BC. No important change would be produced if the 3.8 m value is correlated with the AD 142 event. The earthquake of 1856 contributed proportionally little uplift because of its deep source.

5 *Discussion*

Timing and magnitude of coastal uplifts in NE Rhodes are susceptible to considerable uncertainties. Radiocarbon dating results often provide wide time windows for an uplift episode. The co-seismic 3.8 m uplift near Rhodes city, which was initially correlated with the ~227 BC earthquake (PIRAZZOLI et al. 1989), was found to correspond at the calibrated date window of 178 BC-AD 459 (STIROS & BLACKMAN 2014). The implication is that this time range better matches the AD 142 earthquake. However, other earthquakes are also known to have occurred around that time period, including those of 199-196 BC and AD 115, 344, 474-478, 515. At all evidence these earthquakes have not been as significant as the 227 BC and AD 142 ones, yet some of them fall outside the time window of interest.

The uplift caused by the 1856 earthquake has been documented in the unpublished account of a reliable author. Nevertheless, co-seismic movements caused by sub-crustal earthquakes are rarely reported and, therefore, we investigated for other examples. Relevant instrumental records of recent earthquakes are encouraging, e.g. after the 16 April 2013 southeast Iran earthquake (M_w=7.8) co-seismic offsets were reported from two nearby GPS sites within 300 km of the epicenter (KUNDU et al. 2014). WISEMAN et al. (2012) reported GPS offsets due to the 30 September 2009 Padang (Sumatra) earthquake (M_w=7.6) at epicentral distances up to ~100 km.

At all evidence the 1303, 1741 and 1957 earthquakes didn't cause coastal uplift or subsidence in Rhodes since they had epicenters at estimated distances ranging from 60 to 180 km (Fig. 2), although the 1303 rupture zone may have extended close to Rhodes. According to fault dislocation modeling (KONTOGIANNI et al. 2002, HOWELL et al. 2015) no uplift is expected in NE Rhodes from such distant sources. No surface co-seismic movement was reported from the 1863 and 1926 earthquakes very likely because they have been smaller and deeper as compared to the estimated focal depth (h~60 km) of the 1856 earthquake.

The estimated historical uplift rate of 2.5-3.0 mm a^{-1} is relatively high as compared with the average late Holocene uplift rate of 1 mm a^{-1} (PIRAZZOLI et al. 1989), yet the two estimates are of the same order of magnitude. The higher historical uplift rate may reflect only a short-term phase within the variable long-term uplift rate ranging from 1.4 to 3.9 mm a^{-1} in the last 33 ka (CORNÉE et al. 2018). In the western and south coasts of Crete it ranged from 0.1 to 7.7 mm a^{-1} in the late Quaternary, i.e. the last ~600 ka (TIBERTI et al. 2014, MOUSLOPOULOU et al. 2015, OTT et al. 2019, ROBERTSON et al. 2019). In the west coast of Crete the late Pleistocene-Holocene long-term uplift rate was found 2.5–2.7 mm a^{-1} as an average (TIBERTI et al. 2014).

There is no evidence of permanent uplift accumulating in NE Rhodes predominantly during the inter-seismic period, which is the case also elsewhere (e.g. MELNICK 2016). Aseismic slip is an additional mechanism for uplift in plate margins (e.g. NELSON & MANLEYT 1992). In the HSZ the seismic slip rate represents only a small fraction of plate motion, the rest being accommodated by aseismic slip (e.g. JACKSON & MCKENZIE 1988, PAPADOPOULOS 1989). Aseismic slip, however, may not account for late Holocene uplift in Rhodes, otherwise the late Holocene uplift rate should exceed the historical co-seismic uplift rate. Similarly, continuous sediment underplating does not account for coastal uplift in Rhodes at least during the late Holocene, contrary to findings supporting sustained uplift in Crete due to continuous underplating of sediments (STROBL et al. 2014)

6 *Conclusions*

From field observations we verified the findings of previous publications that the coastal uplift gradient in the east Rhodes increases from 0 at the southern island tip (Prasonisi) to 2.4 m a.m.s.l. at the central segment of coast near Lindos and to 3.8 m a.m.s.l. at the NE tip of the island. We focused our attention at the NE side with the aim to calculate the net, coastal uplift rate during the historical period.

Elevated shorelines and ^{14}C dating indicated that the ~227 BC major (M≥7.7) earthquake was likely associated with coastal uplift of either 3.8 m or 1.5 m. Co-seismic uplift occurred also with the AD ~142 large earthquake, which is possibly matched by ^{14}C dated uplifted shorelines by ~0.3 m to 3.8 m, if the amplitude of 3.8 m is not related to the ~227 BC event. Otherwise the maximum uplift from the AD 142 earthquake was only 1.5 m. Uplift of ~1 m has been dated in a wide time window matching very likely the 18 December 1481 earthquake. The sub-crustal earthquake (h~60 km, M~7.7) of 12 October 1856 caused permanent coastal uplift of ~0.55 m in Rhodes harbor. The earthquakes of 227 BC, AD 142 and 18 December 1481 have been non-tsunamigenic events implying that likely they had their sources close to Rhodes in the shallow water domain. Several significant historical earthquakes (M≥7), such as the 1303, 1741, 1863, 1926 and 1957 ones were incapable to produce coastal uplift in east Rhodes due to source distance or large depth.

The total, net co-seismic uplift in NE Rhodes during the last ~2.3 ka was estimated between 5.75 m and 6.95 m implying coastal uplift rate of 2.5-3.0 mm a^{-1}. This rate falls within the geological uplift range of 1.4-3.9 mm a^{-1} found in the last 33 ka (CORNÉE et al. 2018) but exceeds the late Holocene rate of ~1 mm a^{-1} (PIRAZZOLI et al. 1989). One major earthquake (M~7.7) occurring either in ~227 BC or in AD ~142 released nearly half of the total seismic moment and about 55-65% of the total net, co-seismic historical uplift. The rest uplift was produced by smaller but still large earthquakes (M≥7.0). Coastal uplift due to other mechanisms, e.g. aseismic slip in the plate margin and sediment underplating are not likely at least in the late Holocene period. Our results imply that in NE Rhodes permanent uplift is not predominantly accumulated during the inter-seismic period.

Acknowledgments

The first author was supported by the Hellenic Foundation for Research and Innovation (HFRI) and the General Secretariat for Research and Technology (GSRT), under the HFRI PhD

Fellowship grant (GA. no. 490). We are grateful to Prof. Riccardo Caputo and to one anonymous reviewer for their suggestions that contributed to the substantial improvement of the initial paper submission.

References

Alpar, B., Ünlü, S., Altınok, Y. et al. (2012): New approaches in assessment of tsunami deposits in Dalaman (SW Turkey). - Natural Hazards 63: 181–195, https://doi.org/10.1007/ s11069-010-9692-5.

Ambraseys, N.N. (1962): Data for the investigation of the seismic sea-waves in the Eastern Mediterranean. – Bull. Seism. Soc. Am. 52 : 895-913.

Ambraseys, N.N. (2001): Reassessment of earthquakes, 1900-1999, in the Eastern Mediterranean and the Middle East. - Geophys. J. Int. 145: 471-485.

Ambraseys, N.N. (2009). Earthquakes in the Mediterranean and Middle East, A Multidisciplinary Study of Seismicity up to 1900. Cambridge Univ. Press, Cambridge, UK, 947 pp.

Ambraseys, N. N. & Adams, R. D. (1998): The Rhodes earthquake of 26 June 1926. - J. Seismology 2: 267-292.

Avșar, U. (2019) Sedimentary geochemical evidence of historical tsunamis in the Eastern Mediterranean from Ölüdeniz Lagoon, SW Turkey. - J. Paleolimnol. https://doi.org/10.1007/ s10933-018-00065-x.

Benetatos, C., Kiratzi, A., Papazachos, C. & Karakaisis, G., 2004. Focal mechanisms of shallow and intermediate depth earthquakes along the Hellenic Arc. - J. Geodynamics 37: 253-296.

Blackman, D. J., Knoblauch, P. & Yannikouri, A. (1996): Die Schiffshäuser am Mandrakihafen in Rhodos. - Archäologischer Anzeiger 3: 371-426.

Bocchini, G.M., Brüstle, A. & Becker, D. (2018): Tearing, segmentation, and backstepping subduction in the Aegean: New insights from seismicity. – Tectonophysics: 734–735: 96–118 https://doi.org/10.1016/j.tecto.2018.04.002

Cornée, J.-J., Münch, Ph., Quillévéré, F., et al. (2006): Timing of Late Pliocene to Middle Pleistocene tectonic events in Rhodes (Greece) inferred from magneto-biostratigraphy and 40Ar/39Ar dating of a volcaniclastic layer. - Earth & Plan. Sci. Lett. 250: 281-291.

Cornée, J-J, Quillévéré, F., Moissette, P., et al. (2018): Tectonic motion in oblique subduction forearcs: insights from the revisited Middle and Upper Pleistocene deposits of Rhodes, Greece. - J. Geol. Soc. 176: 78–96, https://doi.org/10.1144/jgs2018-090.

Evelpidou N, Vassilopoulos A. & Pirazzoli P. (2012): Holocene emergence in Euboea island (Greece). - Mar. Geol., 295-298: 14-19.

Flemming, N.C. (1978): Holocene eustatic changes and coastal tectonics in the northeast Mediterranean: implications for models of crustal consumption. - Phil. Trans. R. Soc. A, 289 (1362): 405–458.

Franco, C. (2008): Aelius Aristides and Rhodes: concord and consolidation. In: Harris W.V., et al. (Eds.), Aelius Aristides Between Greece, Rome and Gods, Brill., Leiden: 217-251.

Galanopoulos, A.G. (1960): Tsunamis observed on the coasts of Greece from Antiquity to present time. - Annali di Geofisica 13: 369-386.

Gallen, S.F., Wegmann, K.W., Bohnenstiehl, D.R., Pazzaglia, F.J., Brandon, M.T. & Fassoulas, C. (2014): Active simultaneous uplift and margin-normal extension in a forearc high, Crete, Greece. - Earth and Planetary Sci. Lett. 398: 11-24, https://doi.org/10.1016/j.epsl.2014. 04.038.

Ganas, A. & Parson, T. (2009): Three-dimensional model of Hellenic Arc deformation and origin of the Cretan uplift. – J. Geoph. Res. 114 (B06404), doi:10.1029/2008JB005599.

Gauthier, A. (1979): Contribution a l' étude néotectonique du domaine Egéen: l' île de Rhodes (Dodécanèse,Grèce). - PhD thesis, Academie de Versailles, Université de Paris XI, Centre d'Orsay.

Grünthal, G. & Wahlström, R. (2012): The European-Mediterranean Earthquake Catalogue (EMEC) for the last millennium. – J. of Seismology, 16 (3), 535-570.

Grünthal, G., Wahlström, R. & Stromeyer, D. (2013): The SHARE European Earthquake Catalogue (SHEEC) for the time period 1900-2006 and its comparison to EMEC. J. of Seismology, 17 (4), 1339-1344, doi: 10.1007/s10950-013-9379-y.

Guidoboni, E. & Comastri, A. (2005): Catalogue of earthquakes and tsunamis in the Mediterranean area from the 11th to the 15th century. - INGV-SGA, Rome-Bologna, Italy, 1037 pp.

Guidoboni, E., Comastri, A. & Traina, G. (1994): Catalogue of ancient earthquakes in the Mediterranean area up to the 10th Century. - ING-SGA, Rome-Bologna, Italy, 504 pp.

Guidoboni, E., Ferrari, G., Mariotti, D., Comastri, A., Tarabusi, G., Sgattoni, G. & Valensise, G. (2018): Historical earthquake data from the CFTI5Med catalogue. PANGAEA, https://doi. org/10.1594/ PANGAEA.896754.

Guidoboni, E., Ferrari, G., Tarabusi, G., Sgattoni, G., Comastri, A., Mariotti, D., Ciuccarelli, C., Bianchi, M. G. & Valensise, G. (2019). CFTI5Med, the new release of the catalogue of strong earthquakes in Italy and in the Mediterranean area. Scientific Data, 6(1), https://doi.org/10.1038/s41597-019-0091-9.

Hall, J., Aksu, A.E., Yaltirak, C. & Winsor, J.D. (2009): Structural architecture of the Rhodes Basin: A deep depocentre that evolved since the Pliocene at the junction of Hellenic and Cyprus Arcs, eastern Mediterranean. - Mar. Geol. 258(1–4): 1–23.

Hanken, N.-M., Bromley, R.G. & Miller, G. (1996): Plio-Pleistocene sedimentation in coastal grabens, North-east Rhodes, Greece. - Geol. J. 31 (3): 271-296.

Hanks, T.C. & Kanamori, H. (1979): A moment magnitude scale. - J. Geophys. Res. 84:2348-2350.

Hedenborg, J. (19th century): Geschihte der Insel Rhodos. Autogr. Handschrift (unpublished manuscript).

Howell, A., Jackson, J., England, P. et al. (2015): Late Holocene uplift of Rhodes, Greece: evidence for a large tsunamigenic earthquake and the implications for the tectonics of the eastern Hellenic Trench System. - Geophys. J. Int. 203(1): 459–474.

ISC-GEM (2020): The ISC-GEM Global Instrumental Earthquake Catalogue, Version 7.0, released on 2020-04-09. DOI: http://doi.org/10.31905/D808B825.

Jackson, J. A. & McKenzie, D. (1988): Rates of active deformation in the Aegean Sea and surrounding area. - Basin Res. 1: 121-128.

Kitamura, A., Ina, T., Suzuki, D. et al. (2019): Geologic evidence for coseismic uplift at ~ AD 400 in coastal lowland deposits on the Shimizu Plain, central Japan. – Prog. Earth Planet Sci, 6 (57), https://doi.org/10.1186/s40645-019-0305-y.

Kontogianni, V. A., Tsoulos, N. & Stiros, S.C. (2002): Coastal uplift, earthquakes and active faulting of Rhodes Island (Aegean Arc): modeling based on geodetic inversion. - Mar. Geol. 186: 290-317.

Kundu, B., Nankali, H., Malik, et al. (2014): Coseismic offset due to intermediate depth 16 April 2013 southeast Iran earthquake (Mw7.8). - Bull. Seism. Soc. Am. 104: 1562-1566, Doi: 10.1785/0120130269.

Lambeck, K. (1995): Late Pleistocene and Holocene sea-level change in Greece and southwestern Turkey: a separation of eustatic, isostatic and tectonic contributions. – Goephys. J. Int. 122:1022-1044.

Lekkas, E., Papanikolaou, D. & Sakellariou, D. (1993): Neotectonic map of Greece, 1:100.000 scale, Rhodos. University of Athens, pp. 100.

Løvlie, R., Støle, G. & Spjeldnæs (1989): Magnetic polarity stratigraphy of Pliocene-Pleistocene marine sediments from Rhodos, eastern Mediterranean. - Phys. Earth Planet. Inter. 54: 340-352.

Makropoulos K., Kaviris, G. & Kouskouna, V. (2012). An updated and extended earthquake catalogue for Greece and adjacent areas since 1900. Nat. Hazards Earth Syst. Sci. 12:1425-1430.

Maramai, A., Brizuela, B. & Graziani, L. (2014): The Euro-Mediterranean Tsunami Catalogue. – Ann. Geophys. 57:S0435, doi:10.4401/ag-6437.

McClusky, S., Balassanian, S., Barka, A. et al. (2000): Global Positioning System constraints on plate kinematics and dynamics in the eastern Mediterranean and Caucasus. – J. Geophys. Res.-Solid Earth 105:5695–5719, https://doi.org/10.1029/1999jb900351.

Melnick, D. (2016): Rise of the central Andean coast by earthquakes straddling the Moho. - Nature Geosci. 9:401–407, https://doi.org/10.1038/ngeo2683.

Meulenkamp, J., de Mulder, E. & vad de Werd., A. (1972): Sedimentary history and palaeogeography of the late Cenozoic of the island of Rhodes. - Z. Deutsch. Geo. Ges. 123: 541-553.

Mouslopoulou, V., Nicol, A., Begg, J. et al. (2015): Clusters of megaearthquakes on upper plate faults control the Eastern Mediterranean hazard. - Geophys. Res. Lett. 42, doi:10.1002/2015GL066371.

Mutti, E., Orombelli, G. & Pozzi, R. (1970): Geological studies on the Dodecanese islands (Aegean Sea). X. Geological map of the island of Rhodes (Greece). Explanatory notes. - Ann. Geol. Pays Hell. 22: 77-226.

Nelson, A. R. & Manleyt, W.F. (1992): Holocene Coseismic and Aseismic Uplift of Isla Mocha, South-Central Chile. - Quat. Internat. 15/16: 61-76.

Ott, R. Gallena, F. & Wegmann, K. W. (2019): Pleistocene terrace formation, Quaternary rock uplift rates and geodynamics of the Hellenic Subduction Zone revealed from dating of paleoshorelines on Crete, Greece. - Earth and Planet. Sci. Lett. 525. https://doi.org/10.1016/j.epsl.2019.115757.

Papadopoulos, G. A. (1989): Seismic and volcanic activities and aseismic movements as plate motion components in the Aegean area. - Tectonophysics 167: 31-39.

Papadopoulos, G. A. (2011): A Seismic History of Crete: Earthquakes and Tsunamis, 2000 BC–AD 2010. Ocelotos Publ., Athens, 415 pp.

Papadopoulos, G. A. (2016): Tsunamis in the European-Mediterranean Region: From Historical Record to Risk Mitigation. Elsevier, ISBN: 9870124202245, pp. 271.

Papadopoulos, G. A. & Chalkis, B. (1984): Tsunamis observed in Greece and the surrounding area from antiquity up to the present times. - Mar. Geol. 56: 309-317.

Papadopoulos, G. A. & Lefkopoulos G. (1993): Magnitude-distance relations for liquefaction in soil from earthquakes. - Bull. Seismol. Soc. Am. 83:925-938.

Papadopoulos G. A., Daskalaki, E., Fokaefs, A. & Giraleas, N. (2007): Tsunami Hazard in the Eastern Mediterranean: Strong Earthquakes and Tsunamis in the East Hellenic Arc and Trench System. - Natural Hazards & Earth System Science 7:57–64.

Papadopoulos G. A., Gràcia, E., Urgeles, R. et al. (2014) Historical and pre-historical tsunamis in the Mediterranean and its connected seas: Geological signatures, generation mechanisms and coastal impacts. - Mar. Geol., DOI: 10.1016/j.margeo.2014.04.014.

Papadopoulos, G. A., Minoura, K., Imamura, et al. (2012): Strong earthquakes and tsunamis in the East Hellenic Arc. - Research in Geophysics 2:e12: 90-99, doi:10.4081/rg.2012.e12.

Papathanasiou, G., Pavlides, S., Christaras, B. & Pitilakis, K. (2005): Liquefaction case histories and empirical relations of earthquake magnitude versus distance from the broader Aegean region. - J. Geodynamics 40:275-278.

Papazachos, B. C. & Papazachou, C. (1997): The Earthquakes of Greece. Ziti Publ., Thessaloniki, 304 pp.

Papazachos, B. C. & Papazachou, C. (2003): The Earthquakes of Greece. Ziti Publ., Thessaloniki, 286 pp. (in Greek).

Pavlides, S. & Caputo, R. (2004): Magnitude versus faults' surface parameters: quantitative relationships from the Aegean Region. - Tectonophysics 380: 159-188.

Perrey, A. (1852): Note sur les tremblements de terre ressentis en 1851. - Bull. Acad. R. Sci. Bruxelles 22 (1): 353-396.

Pirazzoli, P. A. (1988): Sea-level changes and crustal movements in the Hellenic Arc (Greece). The contribution of archaeological and historical data. In: A. Raban (Ed.), Archaeology of Coastal Changes, BAR Int. Ser. 404: 157-184.

Pirazzoli, P.A. & Grant, D.R. (1987): Lithospheric deformation deduced from ancient shorelines. In: Kasahara, K. (Ed.), Recent Plate Movements and Deformation, Geodynamics Series v. 2, Intern. Lithosphere Program Publ. No. 0137, Am. Geophys. Union & Geol. Soc. Am.: 67-72.

Pirazzoli, P. A., Thommeret, J., Thommeret, Y., et al. (1982): Crustal block movements from Holocene shorelines: Crete and Antikythira (Greece). – Tectonophysics 86: 27-43.

Pirazzoli, P. A., Laborel, J. & Stiros, S. C. (1996): Earthquake clustering in the easterm Mediterranean during historical times. - J. Geophys. Res. 101: 6083-6097.

Pirazzoli, P. A., Montaggioni, L.F., Saliège, J.F. et al. (1989): Crustal block movements from Holocene shorelines: Rhodes island (Greece) -Tectonophysics 170: 89-114.

Reilinger, R., McClusky, S., Vernant, et al. (2006): GPS constraints on continental deformation in the Africa-Arabia-Eurasia continental collision zone and implications for the dynamics of plate interactions. - J. Geophys. Res. 111, doi:10.10129/ 2005JB004051.

Reilinger, R., McClusky, S., Paradissis, D. et al. (2010): Geodetic constraints on the tectonic evolution of the Aegean region and strain accumulation along the Hellenic subduction zone. - Tectonophysics 488: 22–30.

Robertson, J., Meschis, M., Roberts, G. P. et al. (2019): Temporally constant Quaternary uplift rates and their relationship with extensional upper-plate faults in south Crete (Greece), constrained with 36Cl cosmogenic exposure dating. - Tectonics 38, https://doi.org/10.1029/2018TC005410.

Savage, J. C., Plafker, G., Svarc, J. L. & Lisowski, M. (2014): Continuous uplift near the seaward edge of the Prince William Sound megathrust: Middleton Island, Alaska. - J. Geophys. Res. Solid Earth 119: 6067– 6079, doi:10.1002/2014JB011127.

Schmidt, J. F. (1879): Studien über Erdbeben. - Carl Scholtze, Leipzig, 360 pp.

Skourtsos, E., Pope, R., & Triantaphyllou, M. (2007): Tectono-Sedimentary Evolution and Rates of Tectonic Uplift of the Sfakia Coastal Zone, Southwestern Crete. Bull. Geol. Soc. of Greece: 40(1), 475-487, doi:https://doi.org/10.12681/bgsg.16649.

Sidall, M., Rohling, E. J., Almogi-Labin, A., et al. (2003): Sea-level fluctuations during the last glacial cycle. – Nature 423: 853-858.

Sivan, D., Lambeck, K., Toueg, R. et al. (2004): Ancient coastal wells of Caesarea Maritima, Israel, an indicator for relative sea level changes during the last 2000 years. - Earth Planet. Sci. Lett. 222: 315-330.

Shaw, B., Ambraseys, N. N., England, P. et al. (2008): Eastern Mediterranean tectonics and tsunami hazard inferred from the AD 365 earthquake. - Nature Geoscience, doi:10.1038/ngeo151.

Shaw, B. & Jackson, J. (2010): Earthquake mechanisms and active tectonics of the Hellenic subduction zone. - Geophys. J. Int. 181: 966–984.

Stiros, S. & Blackman, D.J. (2014): Seismic coastal uplift and subsidence in Rhodes Island, Aegean Arc: Evidence from un uplifted ancient harbour. - Tectonophysics 611: 114–120. DOI: 10.1016/j.tecto.2013.11.020

Strobl, M., Hetzel, R., Fassoulas, C. & Kubik, P.W. (2014): A long-term rock uplift rate for eastern Crete and geodynamic implications for the Hellenic subduction zone. - Journal of Geodynamics: 78, 21-31, https://doi.org/10.1016/j.jog.2014.04.002.

Stucchi, M. et al. (2013): The SHARE European Earthquake Catalogue (SHEEC) 1000–1899. - J Seismol. 17:523–544, DOI 10.1007/s10950-012-9335-2.

Taymaz, T., Jackson, J. & McKenzie, D. (1991): Active tectonics of the north and central Aegean Sea. - Geophys. J. Int. 106: 433–490.

ten Veen, J. H. & Kleinspehn, K. L. (2002): Geodynamics along an increasingly curved convergent plate margin: Late Miocene-Pleistocene Rhodes, Greece. – Tectonics 21(3), https://doi.org/10.1029/2001TC001287.

Tiberti, M. M., Basili, R. & Vannoli, P. (2014): Ups and downs in western Crete (Hellenic subduction zone). - Scientific Reports 4:5677, DOI:10.1038/srep05677.

Wiseman, K., Banerjee, P., Bürgmann, R. et al. (2012): Source model of the 2009 Mw7.6 Padang intraslab earthquake and its effect on the Sunda megathrust.- Geophys. J. Int. 190: 1710–1722, doi:10.1111/j.1365-246X.2012.05600.x.

Wells, D.L. & Coppersmith, K.J. (1994): New empirical relationships among magnitude, rupture length, rupture width, rupture area, surface displacement. - Bull. Seismol. Soc. Am. 84: 974–1002.

Woodside, J., Mascle, J., Huguen, C. & Volkonskaïa, A. (2000): The Rhodes Basin, a post-Miocene tectonic trough. - Mar. Geol. 165(1–4): 1–12.