

Chapter 2

The AD 365 Earthquake: Large Tsunamigenic Earthquakes in the Hellenic Trench

...and solid earth was shaken and trembled, the sea with its rolling waves was driven back and withdrew from the land ... Hence, many ships were stranded as if on dry land, and since many men roamed without fear in the little that remained of the waters, to gather fish and similar things with their hands, the roaring sea, resenting this forced retreat, rose in its turn; and over the boiling shoals it dashed mightily upon islands and broad stretches of the mainland, and levelled innumerable buildings in the cities and wherever else they are to be found; so that amid the mad discord of the elements the altered face of the earth revealed marvellous sights. For the great mass of waters, returning when it was least expected, killed many thousands of men by drowning; and by the swift recoil of the eddying tides a number of ships, after the swelling of the wet element subsided, were found to have been destroyed, and the lifeless bodies of shipwrecked persons lay floating on their backs or on their faces. Other great ships, driven by the mad blasts, landed on the tops of buildings (as happened at Alexandria), and some were driven almost two miles inland, like a Laconian ship which I myself saw in passing that way near the town of Motho [Mothoni/Methoni], yawning apart through long decay.

Ammianus Marcellinus, writing after AD 378.

2.1 Summary

Historical accounts describe an earthquake and tsunami on 21 July AD 365 that destroyed cities and drowned thousands of people in coastal regions from the Nile delta to modern-day Dubrovnik. The location and tectonic setting of this earthquake have been uncertain until now. In this chapter, I present evidence from radiocarbon data and field observations that western Crete was lifted above sea-level, by up to 10 m, synchronously with the AD 365 earthquake. The distribution of uplift,

combined with observations of present-day seismicity, suggest that this earthquake occurred not on the subduction interface beneath Crete, but on a fault dipping at about 30° within the overriding plate. Calculations of tsunami propagation, carried out by Piggott and colleagues, show that the uplift of the sea floor associated with such an earthquake would have generated a damaging tsunami through much of the eastern Mediterranean. GPS measurement of the present rate of shortening between Crete and the Cyclades can be used to estimate an approximate repeat time of ~ 5000 years for tsunamigenic events on this single fault in western Crete. If the same process takes place along the entire Hellenic subduction zone, such events may occur approximately once every 800 years.

2.2 Introduction

The AD 365 earthquake was felt throughout the eastern Mediterranean, and its tsunami inundated coastal sites in Africa, the Adriatic, Greece and Sicily. Figure 2.1 shows sites where earthquake damage was recorded in the historical record (yellow dots), and the sites that are known to have been inundated by the tsunami (blue triangles). Rich agricultural land in the Nile delta was abandoned as a result of the flooding, and previously populous hill cities were subsequently inhabited only by hermits [1].

The cause of the AD 365 event, and of other large earthquakes and tsunamis in the Eastern Mediterranean, such as those in AD 551 [2] and AD 1303 [3], has been enigmatic. Most of the world's devastating tsunamis are thought to be generated by slip on the interfaces between plates at subduction zones. The Hellenic subduction zone is the only such major plate boundary in the region, but appears to be largely decoupled; just 10% of the total Nubia-Aegean convergence has been accommodated in earthquakes in the last 100 years (see Chap. 3 for summary).

The key to understanding the kinematics of the area lies in a paleo-shoreline that fringes the coasts of western Crete and Antikythera. This shoreline was first described by Captain Thomas Spratt, RN, in 1851 [4]. Spratt noted many 'sea marks' around the coast of western Crete, reaching a maximum elevation of 10m above present sea-level in the south-west corner of the island. Figure 2.2 shows some examples of his sketches, with modern-day field photographs for comparison. Because the top of the marks runs through the remains of a Roman harbour at Phalasarna, which is now 6m above sea-level, he deduced that the land must have been raised during or after the Roman era (Fig. 2.2). Pirazzoli et al. [5] showed that this shoreline had a ^{14}C age of around 2,000 years BP and attributed its uplift to an earthquake; this earthquake was subsequently linked to the AD 365 event [6–8].

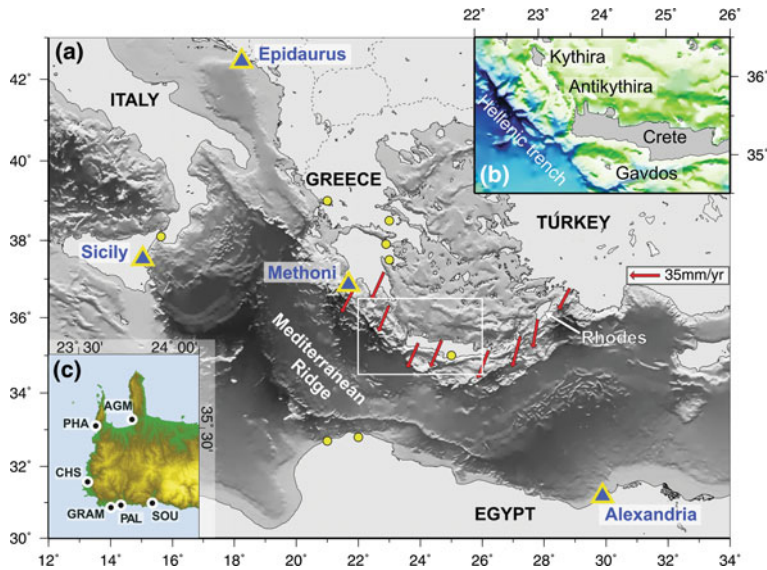


Fig. 2.1 **a** Yellow dots mark places where the AD 365 earthquake is recorded to have caused destruction, blue triangles mark places known to be affected by the associated tsunami. Red arrows show velocities of GPS stations near the Hellenic Trench relative to the Nubian plate from McClusky [40], McClusky et al. [41] and Reilinger et al. [42]. **b** Principal localities referred to in this chapter; note the prominent NW-SE bathymetric escarpment of the Hellenic Trench. **c** Locations of samples whose radiocarbon dates are reported in Table 2.1

2.3 Recent Surface Uplift in Western Crete and Antikythira

The starting point for this chapter is to establish whether, as suggested previously, all of the recent uplift in western Crete indeed took place in one event, close to AD 365. In what follows, the term ‘uplift’ is used to denote upward displacement of rocks relative to the geoid [9]. Because global sea-level has been stable over the interval under consideration, it is possible to take the present elevation of the paleo-shoreline as a proxy for such uplift.

A single prominent paleo-shoreline can be followed around the whole of western Crete; the algal encrustations marking this shoreline are not seen anywhere on the cliffs below this level, but similar constructions can be seen growing at sea-level today (Fig. 2.3). The elevation of the paleo-shoreline decreases from the south-western corner of Crete. Where the shoreline is highest, it is marked by a thick white algal encrustation (Fig. 2.3), but where the shoreline is lower (1–2 m above present day sea-level) it is marked by a deep erosive notch (Fig. 2.4). Previous radiocarbon dates of material from the upper levels of the paleo-shoreline were determined largely by conventional (non-AMS) methods and gave calibrated ages that lie within ~200 years of AD 365 [7, 10]. Taken together, these observations suggest that some uplift took place around AD 365, but they do not give direct evidence that the whole 10 m of

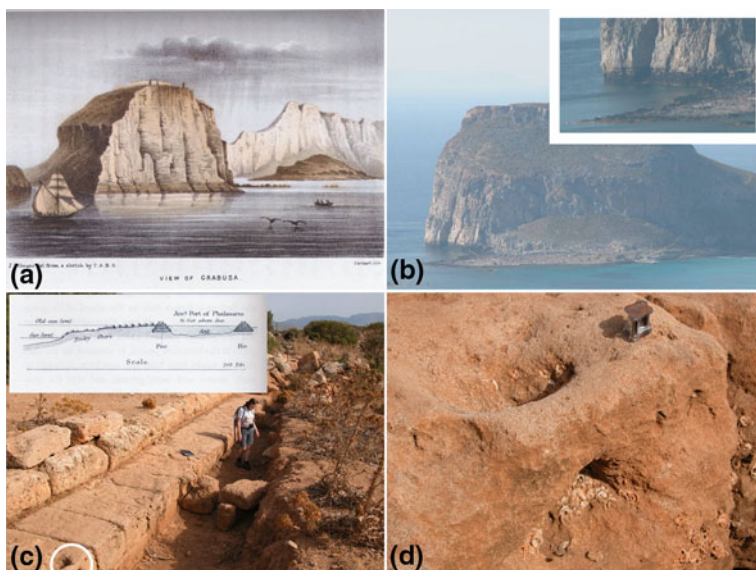


Fig. 2.2 Evidence of uplift of the western part of Crete, described by Thomas Spratt, R.N. **a** Gramvousa peninsula from an engraving by Spratt. **b** The same peninsula photographed today from $35^{\circ} 35.244'N$, $23^{\circ} 35.682'E$. The AD 365 shoreline can be recognised as a dark band around the base of the cliffs in both the engraving and the photograph (shown more clearly in the close-up *inset*). **c** The uplifted harbour at Phalasarna ($35^{\circ} 30.593'N$, $23^{\circ} 34.115'E$). Encrustations on the harbour wall reveal the position of sea-level when the harbour was operational; this level is now 6 m above sea-level. The *inset* is a sketch by Spratt showing the position of the harbour walls relative to sea-level. **d** Details of an encrusted stone ring, *circled* in (c), that was used to tie up mooring ships

uplift took place in one step, and the attribution of the uplift to the AD 365 event was made uncertain by issues related to the calibration of the radiocarbon ages [11]. Here, I address both those sources of uncertainty with new, more accurate, radiocarbon dates on material from elevations between present sea-level and the uplifted paleo-shoreline.

Corals, bryozoans, and *Lithophaga* were collected from widely separated locations in western Crete, at elevations between 1.5 m above sea-level and the height of the paleo-shoreline (Figs. 2.5, 2.6). At all localities visited, the most recent colonisers appeared to be cheliostome bryozoans and ahermatypic corals, of which several had grown inside *Lithophaga* holes. ^{14}C dates were obtained for 13 of these corals and bryozoans. For nine of them, calibrated accelerator mass spectrometry (AMS) ^{14}C dates have 1σ ranges (typically $\sim \pm 70$ years) that include AD 365, and all but two of them include AD 365 in their 2σ age ranges (typically $\sim \pm 140$ years) (Table 2.1). All the *Lithophaga* shells gave dates older than AD 365 (see Chap. 4 for discussion).

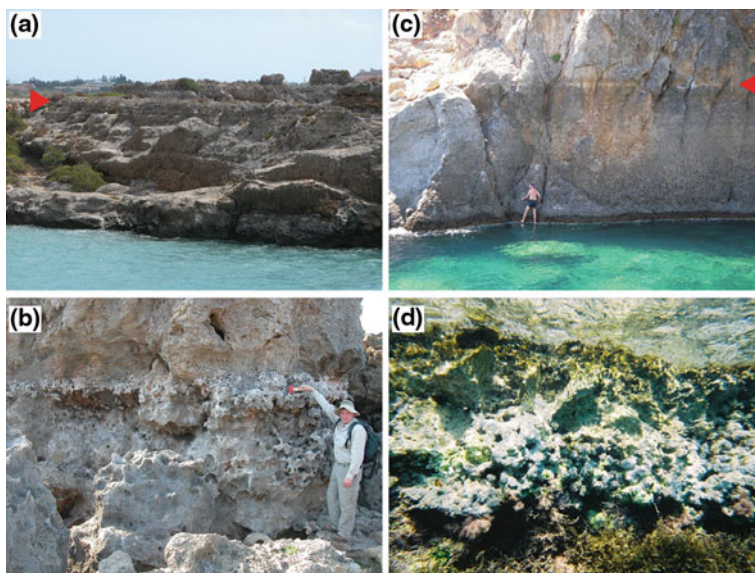


Fig. 2.3 Clear algal encrustation marking the AD 365 shoreline. **a** At Chrisoskalitissa ($35^{\circ} 18.857'N$, $23^{\circ} 32.008'E$) the top of the shoreline reaches an elevation of 8 m. **b** Close-up of the preserved algal encrustation, again from Chrisoskalitissa. **c** In the harbour at Soughia ($35^{\circ} 14.755'N$, $23^{\circ} 49.275'E$) the shoreline can be recognised at 6.7 m above sea-level. **d** A similar encrustation growing at sea-level today [the photograph was taken under water in the bay to the east of the harbour at Soughia ($35^{\circ} 14.864'N$, $23^{\circ} 49.078'E$)]

Before treatment, I examined the samples under an SEM to confirm the presence of primary structural fabrics (Figs. 2.5, 2.6) and selected the cleanest pieces for dating. I thoroughly cleaned all but four coral and bryozoan samples by shot blasting. Samples Phalasarna 1, Paleochora 8, Phalasarna 3 and Soughia 2 were instead cleaned ultrasonically in distilled water before being submerged in a bath of dilute (0.1 M) hydrogen peroxide until effervescence ceased.

Carbonate samples were dated using AMS ^{14}C dating by the Oxford Radiocarbon Accelerator Unit (ORAU), University of Oxford (see Chap. 4 for details). Radiocarbon determinations were calibrated using the INTCAL04 Marine Calibration dataset [12] and the OxCal computer calibration programme. A ΔR value of 53 ± 43 years was used to account for the local reservoir offset (R) from the modelled world ocean [13]. This value is the average for the Eastern Mediterranean region. The calibrated data are shown in Table 2.1. The bryozoan sample Chrisoskalitissa 5 was analysed twice (17671, 17672, Table 2.1) to check reproducibility.

These new radiocarbon dates require that almost all of the uplift of western Crete took place within a few decades, i.e. within the uncertainties of dating, of AD 365. (I have no datable material from locations lower than 1.5 m above sea-level.) Although the data permit slow uplift during that interval, or uplift in a series of rapid small events, uplift of this magnitude in a single earthquake would have caused

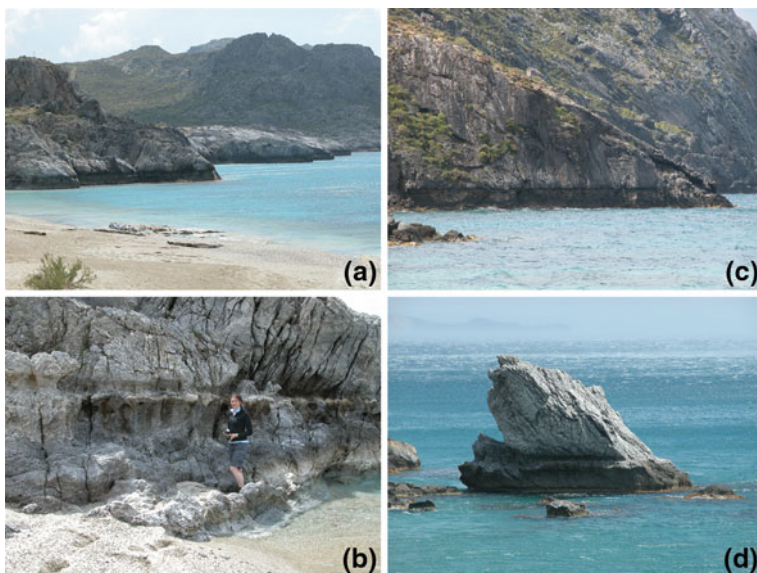


Fig. 2.4 Erosive notch (no algal encrustation) marking the AD 365 shoreline. The top of the shoreline is relatively low (between 1 m and 2 m above present day sea-level). **a, b** Damiani ($35^{\circ} 10.49'N$, $24^{\circ} 24.86'E$) at the eastern and western sides of the bay; **c** Plakias ($35^{\circ} 10.98'N$, $24^{\circ} 24.07'E$); **d** 1 km east of Amondi Bay ($35^{\circ} 09.19'N$, $24^{\circ} 28.86'E$)

the historically recorded tsunami of AD 365 (see below). Therefore either the uplift of western Crete and its surrounding seafloor took place slowly within a few decades of AD 365 and some other event caused the tsunami that destroyed Alexandria in AD 365, or more likely the two events are connected.

2.4 Fault Slip During the AD 365 Earthquake

Measurements of ground displacements after an earthquake are commonly used to place bounds on the orientation and location of the fault that slipped. In the present case, such measurements are limited to uplift, in the sense defined above, of paleo-shorelines, and I use information from the distribution of earthquakes in the region as additional constraints on the location of the fault that caused the AD 365 earthquake.

The most obvious candidate fault is the subduction interface, whose orientation and location are revealed by the focal mechanisms of earthquakes striking NW-SE at 40 km depth on a shallowly-dipping ($\sim 15\text{--}20^{\circ}$) plane, directly below the SW corner of Crete [14], (Chap. 3, Fig. 2.7). This interface projects to the surface about 150 km SW of Crete, where the Mediterranean Ridge, a prism of sediments ~ 10 km thick, is deforming aseismically by folding and thrusting [15]. The interface appears to be

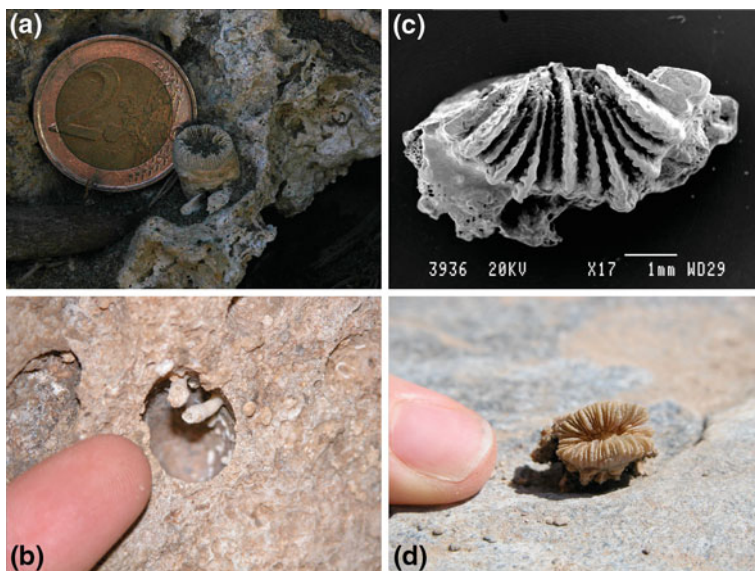


Fig. 2.5 **a** *Balanophyllia regia*, collected at Paleochora; **b** *Stenocyathus vermiformis*, collected at Phalasarna from within *Lithophaga* holes; **c** *Caryophyllida*, collected at Aghia Marina and imaged using SEM; **d** *Caryophyllida*, collected at Aghia Marina

aseismic beneath 40–45 km depth, in common with other subduction zones around the world [16].

Figure 2.8a–c is a cartoon showing the pattern of surface deformation during a seismic cycle in a typical subduction zone. After an earthquake, the interface is locked (a) and the overriding plate is dragged down (b). Points at the surface that are above the locked zone are dragged down, whereas points further back (above the aseismic and freely slipping deeper part of the interface) are uplifted in the flexural bulge. During an earthquake (c) the vertical motion is suddenly reversed: the region above the seismogenic part of the interface is uplifted, and points landward of this subside. Moderate-sized earthquakes on the subduction zone interface have a maximum depth of 40–45 km in this part of the Hellenic arc, and events at this depth lie directly beneath the south-west corner of Crete (Fig. 2.7, Chap. 3). If the subduction zone interface slipped between the surface and 40 km depth, Crete would lie on the landward side of the base of the seismogenic part of the interface, and would therefore be expected to subside in such an earthquake.

Figure 2.8d–e shows the calculated elastic surface deformation from an earthquake modelled as slip along the subduction zone interface between the surface and 40 km depth, using a planar dislocation in an elastic half-space [17]. The interface is assumed to dip at 15° and projects to the surface 150 km to the south-west of Crete (beneath the Mediterranean Ridge). As expected, Crete is downthrown, rather than uplifted by this earthquake, as the deepest extent of slip is seaward of the island. The

Table 2.1 ¹⁴C dates of samples from western Crete (identified by Oxford Radiocarbon Accelerator Unit sample number). Age ranges in bold include AD 365

Ox ID	Organism	Location	Height (m)	Top of paleo-shore (m)	¹⁴ C age BP	1σ range Years	2σ range Years
16743	Coral (S)	PHA 1	3.5	6.6	1985 ± 28	AD 415–546	AD 340–604
16745	Coral (B)	PAL 8	9.0	9.0	2019 ± 28	AD 360–507	AD 282–565
16991	Coral (S)	PHA 3	3.0	6.6	2032 ± 28	AD 338–484	AD 271–551
17669	Bryozoans (My)	CHS 1	3.3	7.9	2129 ± 27	AD 238–378	AD 159–428
17670	Bryozoans (My)	CHS 2	3.0	7.9	2071 ± 27	AD 297–431	AD 235–515
17671	Bryozoans (My)	CHS 5	2.5	7.9	2019 ± 27	AD 360–506	AD 284–564
17672	Bryozoans (My)	CHS 5	2.5	7.9	2024 ± 26	AD 351–500	AD 280–556
17673	Coral (Ca)	CHS 7	6.0	7.9	2345 ± 27	BC 22–AD 109	BC 100–AD 179
17674	Coral (Ca)	AGM	3.0	5.5	1977 ± 26	AD 423–550	AD 352–608
17675	Bryozoans (My)	SOUNB2	2.2	6.7	2119 ± 29	AD 249–388	AD 166–438
17676	Bryozoans (My)	GRAM2	4.7	9.0	1878 ± 25	AD 539–653	AD 456–675
17677	Bryozoans (My)	GRAM4	2.0	9.0	2015 ± 28	AD 366–514	AD 288–570
17678	Coral (S)	PHA 2	3.4	6.6	2028 ± 27	AD 344–489	AD 275–553
13809	Bivalve	CHS T1	20–24	7.9	39180 ± 300	41101–40661	41851–40327
14085	Bivalve	CHS T2	20–24	7.9	37700 ± 250	40615–39981	40649–39237
16995	Bivalve	PAL T	20–24	9.0	47300 ± 550	53063–45967	53461–45621

Coral and bryozoans species were taken from or below the uplifted paleo-shoreline: *S Stenocyathus vermiformis*, *B Balanophyllia regia*, *Ca Caryophyllida*, *My Myriapora truncata*. The bivalves were taken from marine terraces at 20–24 m above sea-level, and at least 10 m above the paleo-shoreline associated with the AD 365 event. Abbreviated location names correspond to labels in inset on Fig. 2.1 : *AGM* Aghia Marina, *PHA* Phalasarna, *CHS* Chrysokallitissa, *GRAM* Gramena, *PAL* Paleochora, *SOU* Soughia. The 1σ and 2σ ages for samples 13809, 14085 and 16995 are years BC, and were obtained by comparing the radiocarbon ages with the ¹⁴C record from the Cariaco Basin [45] with a 400 year subtraction made from the conventional ages to account for the marine reservoir effect

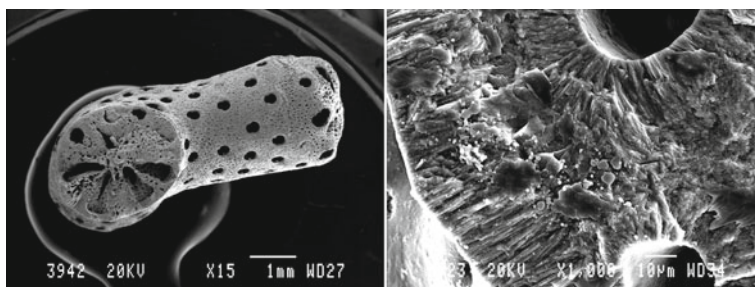


Fig. 2.6 Example of a bryozoan sample (*Myriapora truncata*) imaged using the SEM. The *first* image shows the entire organism, the *second* is a close-up of the crystalline structure. The aragonite needles radiate from the pores, and from the edge of the organism, showing that the primary crystalline fabric has been preserved

only way to force the uplift of Crete by slip along the subduction zone interface is to have coseismic slip down to depths of ~ 70 km, far deeper than the seismogenic thickness of the region. Slip along the subduction zone interface therefore does not seem to be a plausible mechanism for producing the AD 365 uplift.

The distribution of shallow earthquakes (Fig. 2.7) suggests that other faults may be active in the region between the Mediterranean Ridge and the coast of Crete. The most prominent surface feature in this area is the Hellenic Trench, a linear NW-SE escarpment 3,500 m deep at its lowest point, 25 km SW of Crete (Fig. 2.7). The proximity of this escarpment to the highest ground on Crete, where there is abundant geomorphological evidence of recent uplift (Chap. 5), suggests that it marks the outcrop of a major reverse fault dipping beneath Crete. The Hellenic Trench strikes at 315° , parallel to the Mediterranean Ridge and perpendicular to the slip-vectors of thrust-faulting earthquakes in this region. In order to determine possible causative faults for the AD 365 earthquake, I searched for faults that best fit the uplift data from western Crete. I restricted candidate faults for the AD 365 earthquake to faults with this strike and cropping out between the Mediterranean Ridge and the coast of Crete. This range bounds the plausible locations for reverse faults that could cause the observed uplift; a fault below the subduction interface is unlikely, a reverse fault cropping out on land would drop the south-western-most part of Crete, not raise it. I systematically searched this parameter space to find the fault that yielded the best fit (in the least-squares sense) to the observed uplift.

I modelled slip on candidate faults as planar dislocations buried within an elastic half-space [17]. I assumed that relative motion was in the dip-slip sense, as seen in modern fault-plane solutions for the region; the uplift data, in any case, do not constrain the strike-slip component of motion. I assumed that the fault slipped from the surface to a depth of 45 km, a maximum possible depth suggested by the hypocentres of micro-earthquakes within the overriding plate, which are all shallower than 45 km [18, 19], and by the maximum depths of shallowly dipping thrust faulting on the plate interface [14], Fig. 2.7 and Chap. 3. With these assumptions, the free parameters of

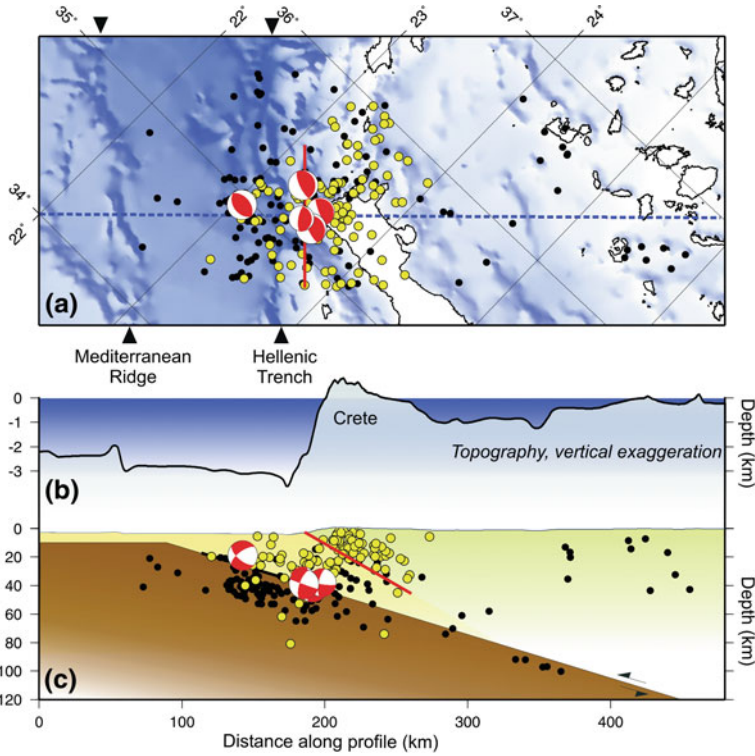


Fig. 2.7 Seismicity and topography in the area of Crete. **a** Seismicity of the region corresponding to the AD 365 earthquake. *Yellow dots* show epicentres of earthquakes determined from teleseismic data [42], *black dots* depict epicentres of microearthquakes determined from local networks [18, 43]. Fault-plane solutions for earthquakes with well-determined depths on the plate interface are from Taymaz et al. [14] and from this study. The projection to the surface of the best-fitting fault for AD 365 earthquake (Fig. 2.9) is shown by *red line*. **b** Vertically exaggerated topography showing the locations of the Mediterranean Ridge and Hellenic Trench. **c** Topography and seismicity projected onto a vertical section along *dotted line* in the centre of panel (a), with our interpretation of the structure. The plate interface is shown by a *black line* and focal mechanisms, the best-fitting fault (Fig. 2.9) by a *red line*. Symbols for hypocentres in (c) as for epicentres in (a). The *brown region* represents the subducting African lithosphere and the *green region* is the overriding Aegean

the candidate faults are their dips, the distance from Crete that they project to the surface, and the amount of slip. The end points of the fault are constrained by uplift of 2.7 m on Antikythira in the NW (of similar appearance and age to that on Crete, [20, 7]), and by my observation that there is no uplifted paleo-shoreline on the island of Gavdos, in the south-east, so the length of the fault was set at 100 km.

The best fit to the uplift data is obtained for a fault that crops out near the Hellenic Trench (Fig. 2.9). The fault length is 100 km, the slip is 20 ± 5 m to a depth of 45 km, the dip is $30 \pm 5^\circ$, and the RMS misfit to the observations is 0.8 m (Fig. 2.9). If the causative fault is forced to have a dip as shallow ($\sim 15^\circ$) as that of the plate interface,

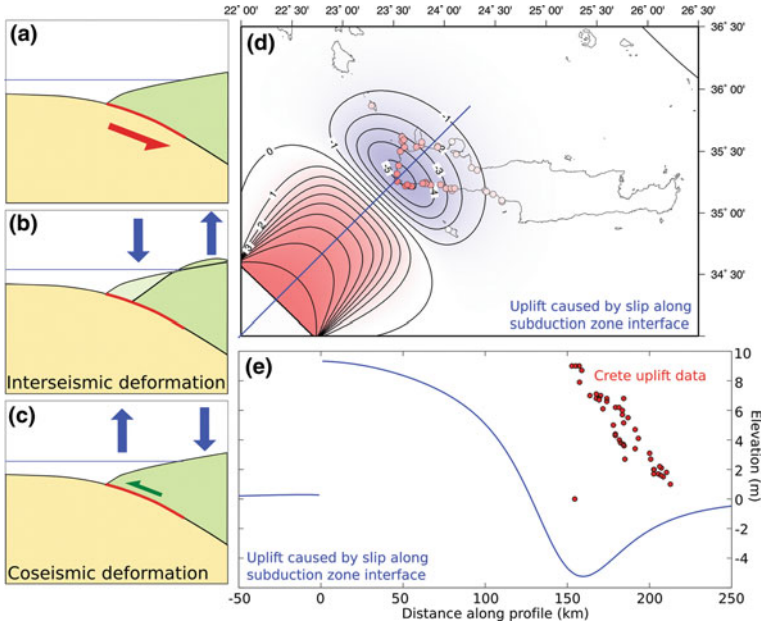


Fig. 2.8 a–c Cartoons showing the pattern of surface deformation during a seismic cycle in a typical subduction zone, modified from a cartoon by James Jackson. After an earthquake, the interface is locked (a) and the overriding plate is dragged down (b). Points seaward of the ‘hinge’ (which lies above the base of the locked zone) are dragged down, whereas points further back (above the aseismic and freely slipping part of the interface) are uplifted in the flexural bulge. During an earthquake (c) the vertical motion is suddenly reversed: the region above the seismogenic part of the interface are uplifted, and points landward of the ‘hinge’ subside. **d** Pattern of uplift expected if slip occurs along the subduction zone interface, between the surface and 45 km depth. Contours are plotted every metre, and the surface is shaded according to uplift. The island of Crete should subside, rather than being uplifted. The pink dots show the elevation of the AD 365 shoreline around the coast of Western Crete. The blue line shows the surface trace of the cross-section. **e** Cross-section of the surface deformation from the interface-slip model in (d), plotted as elevation (m) against distance along the profile. The AD 365 shoreline data are plotted as red dots

but the distance from Crete and the slip are allowed to vary, then the best-fitting solution, with an RMS misfit of 0.9 m, yields a fault that crops out only 70–80 km SW of Crete, where there is no obvious topographic expression of faulting. Any fault dipping at $\sim 15^\circ$ requires slip of 45 ± 10 m to match the observed uplift, which is greater than has been recorded for any earthquake. The trade-offs between parameters are illustrated in Fig. 2.10. The red dot in each of the plots represents the global minimum (i.e. the solution that best fits the measured uplift data). Contours show increasing misfit (in metres) between the model and the data. If the fault is constrained to have a shallow dip, more slip must occur to produce the observed uplift (Fig. 2.10b). The blue dot shows the best-fitting solution if the dip of the fault is constrained to be 15° .

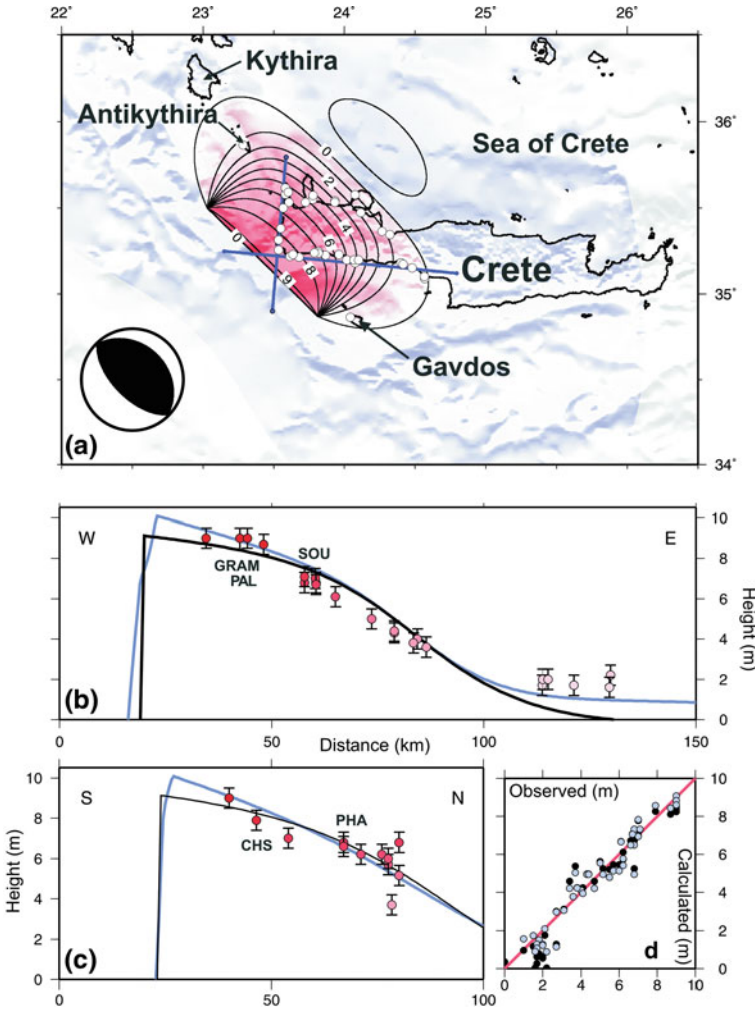


Fig. 2.9 Observed and modelled uplift associated with the AD 365 earthquake. **a** Contours of uplift calculated from the best-fitting elastic dislocation fault model, represented by the *focal mechanism*. *White circles* indicate sites of uplift observations on Crete, Antikythira and Gavdos. **b** W-E profile of shoreline elevations along south coast of Crete measured by Pirazzoli et al. [7], Spratt [4] and ourselves, with 1 m error bars representing uncertainties arising from tidal range and measurement errors. The *black line* shows uplift calculated for best-fitting elastic model for earthquake; the *blue line* shows calculation allowing for post-seismic viscous relaxation (see text). **c** As for (b) but for S-N profile along west coast. **d** Observed and modelled elevations of paleo-shorelines for the bestfitting elastic dislocation model, with 20 m of slip (*black circles*) and for 25 m of slip, followed by complete post-seismic relaxation (*blue circles*)

A previously published model of the uplift [21] required a fault dipping at 40° to a depth of 70 km, well below the maximum seismogenic depth of 45 km beneath Crete. I have found solutions that allow slip to such a depth, but regard them as unlikely

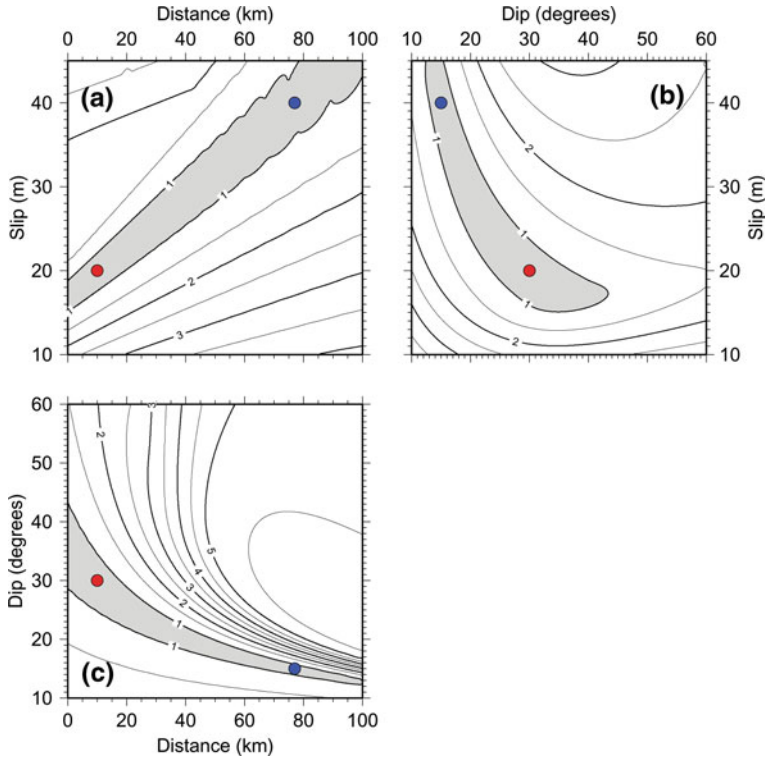


Fig. 2.10 RMS misfit (in metres) between measured and modelled shoreline heights, showing the trade-offs between: **a** slip and the distance of the fault outcrop from the SW corner of Crete; **b** dip of and slip on the fault; **c** dip and the distance of the fault outcrop from the SW corner of Crete. Red dots mark the best-fitting solution and blue dots show the best-fitting solution for a fault with dip fixed to 15°

to represent what occurred in the AD 365 earthquake, because they are generated only by the need to fit uplift of ~ 2 m beyond 100 km from the SW corner of Crete (Fig. 2.9b). Post-seismic relaxation after a thrust-faulting earthquake tends to cause uplift above points landward of the base of the co-seismic fault, and I regard it as more likely that the distant uplift is caused by post-seismic relaxation than by deep co-seismic slip. The data do not warrant a sophisticated analysis of post-seismic relaxation, but I have tested the plausibility of this suggestion by a combined co- and post-seismic calculation [22] in which the fault cuts through an upper, purely elastic, layer of 45 km thickness, that lies upon a visco-elastic half-space. This calculation shows that 25 m of slip on the best-fitting fault, followed by complete viscous relaxation (for which case the viscosity of the half-space need not be specified), accounts better for the observations of uplift than does the purely elastic model, particularly in its ability to fit the distant uplift of ~ 2 m (Fig. 2.9d) (25 m, rather than 20 m, co-seismic slip is required because viscous relaxation causes subsidence close to the fault on the side of the fault that was uplifted in the earthquake). Intriguingly, where

uplift is less than 2 m, the algal encrustation that marked the shoreline at higher elevations is replaced by a deep erosive notch (compare Fig. 2.3 with Fig. 2.4). If the uplift here occurred gradually as post-seismic, rather than co-seismic uplift, the encrustation may have been removed by the wave-action.

The parameters of the best-fitting dislocation solution allow me to estimate the minimum seismic moment for the AD 365 event. These parameters are equivalent to an earthquake of magnitude M_w 8.3–8.5. The ratio of slip (~ 20 m) to length (~ 100 km) for this fault is larger than normally observed [23], but the fault is similar in area and displacement to one of the maximum-slip fault patches in the M_w 9.3 Sumatra earthquake of 2004 [24], and to the 1897 Assam earthquake [25].

2.5 The Tsunami

The tsunami simulations carried out here were performed with ICOM (the Imperial College Ocean Model, <http://amcg.ese.ic.ac.uk/ICOM> [26]) by Piggott and colleagues. ICOM discretises the Navier-Stokes equations in 3D and assumes the Boussinesq approximation. Finite element discretisation techniques are employed on unstructured meshes, with the discretisation and mesh formed in Cartesian coordinates on the surface of an assumed sphere of radius 6,378 km. The starting point for the tsunami simulation is the construction of an unstructured mesh representing the domain of interest. One minute GEBCO bathymetric data of the entire present-day Mediterranean Sea was optimised using Terreno (<http://amcg.ese.ic.ac.uk/Terreno>, [27]); the result is a one-element-deep tetrahedral mesh of approximately 2.8×10^5 nodes and 8.2×10^5 elements. No wetting/drying capability was used here, that is, the chaotic run-up that occurs when the tsunami-wave oversteepens and breaks was not modelled, so the mesh stops at the 20 m bathymetric depth contour. A numerical ‘tide gauge’ was placed at the edge of the mesh approximately 3 km off-shore of Pharos (near Alexandria, Egypt). Crank–Nicolson time-stepping was used with a time-step of 30 s. Continuous piecewise-linear basis functions are used for the spatial discretisation with a second-order anisotropic Smagorinsky LES turbulence model where the filter length scale depends on the local mesh size and shape.

The tsunami calculations were carried out using the best-fitting model of fault slip to provide initial conditions for the submarine surface displacement. Figure 2.11 shows snapshots of the calculated sea surface height at times of 4–90 min after the earthquake, using the surface displacement pattern shown in Fig. 2.9 as the initial condition. The calculation shows that direct waves have the most significant effect, travelling towards the Nile Delta and up the Adriatic. The phases reflected off shorelines interfere and become incoherent in the relatively enclosed Mediterranean Basin.

A prominent feature of the calculation is the wave sweeping east along the African coast to the Nile delta (Fig. 2.11). In AD 365 the offshore island of Pharos (site of the famous lighthouse) was linked to the land by a causeway, the Heptastadion, which was vulnerable to the SW and flooded by the tsunami. It was here that much destruction was concentrated.

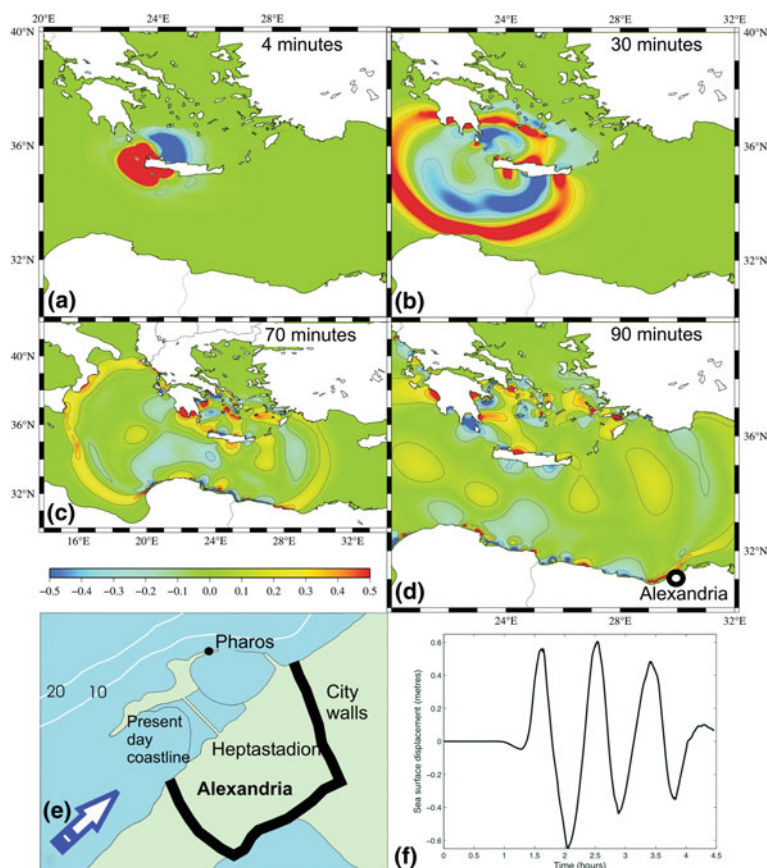


Fig. 2.11 Tsunami calculation. **a–d** Heights of the sea surface at 4, 30, 70 and 90 min after the earthquake. The direct wave travels to the north African coast and into the Adriatic, consistent with the historical record. **e** Sketch map of Alexandria at the time of the AD 365 earthquake. The island of Pharos was connected to the mainland by a narrow causeway, known as the Heptastadion, which was overtopped by the tsunami. Our calculations show that the wave would have arrived from the SW (white arrow), explaining how the Heptastadion was inundated despite the apparent protection offered by Pharos Island against a direct wave from Crete (which lies to the NW). **f** Wave height against time that would be observed in the open ocean off Alexandria in water of 20 m depth. The wave height would have been greatly amplified during run-up through shallower water towards the city

Calculation of the sea surface height at a numerical ‘tide gauge’ situated in the mesh just off the shore of Alexandria shows wave heights of ± 0.6 m in the open ocean (Fig. 2.11f). There are many non-linear effects of the detailed near-shore bathymetry that make it difficult to convert this range into an estimate of the run-up in ancient Alexandria, not least because local conditions of bathymetry and land surface have changed since AD 365 (especially on the Nile Delta). However, the open-ocean

amplitude of the wave is comparable with that observed and modelled in the open ocean for the 2004 Sumatra tsunami [28], and therefore the on-shore effects of such a sea wave presumably would be devastating.

A simulation was also run to calculate the tsunami expected from 40 m of slip on the fault dipping at 15°; the essential characteristics of this tsunami are very similar to those generated by the best-fitting fault, differing principally in that wave heights are greater by a factor of up to five. Even such large differences cannot reliably be resolved from the documented historical evidence and either modelled tsunami is compatible with what is known of the AD 365 inundation, with the direct wave reaching all of the coastal sites described in historical records from the Nile delta and north Africa to Sicily and the Adriatic, all of which are much more heavily populated now than at the time of the earthquake.

2.6 Which is the Tsunamigenic Fault Beneath Crete?

Assessment of the significance of the AD 365 earthquake hinges on understanding its role in accommodating the ~ 35 mm/year of convergence between African and Eurasian lithosphere across the Hellenic subduction zone. Over the past 100 years (in which the record of earthquakes is, for this purpose, complete), earthquakes can account for no more than $\sim 10\%$ of that convergence (see Sect. 3.2.1), so it is necessary to determine whether that percentage is likely to be representative of the longer term, or whether a significant fraction of the convergence could be accommodated by rare, very large, earthquakes.

If the long-term convergence took place principally in large earthquakes like the AD 365 event, then, in SW Crete, such events would repeat approximately every 550 years (for 20 m slip on the steeper fault) or every 1,100 years (for 40 m slip on the plate interface). But field observations and radiocarbon dates show that the 2,000-year-old paleo-shoreline was lifted close to its present position within a few decades of the AD 365 earthquake, implying that this was the *only* earthquake of significant size in that region in the past 1,650 years. Additionally, where the paleo-shoreline is preserved, there is no evidence for previous AD 365-type events since sea-level stabilised near its present height about 6,000 year BP.

This argument can be extended to the whole Hellenic subduction zone, which is about 600 km long. To accommodate convergence across the whole zone by seismic slip in earthquakes would require an earthquake equivalent to the AD 365 event every 100–200 years. The historical record shows a small number of earthquakes in this region (2 or 3 in the past 2,000 years) that could have been of such magnitude but, although the record is incomplete, it is inconceivable that the 10–20 or more great earthquakes that would be needed to remove the seismic slip deficit could have occurred unremarked. For these reasons, I conclude that most of the plate convergence in this region is accommodated aseismically, as did Jackson and McKenzie [29]. This conclusion is reinforced by present-day GPS measurements described later, which show that about 90% of the convergence rate is occurring by steady slip.

If there is some locking along a fault to the south-west of Crete, lines between the Cyclades and Crete should shorten with time. The Cyclades are moving SW at almost the same velocity as Crete, showing that very little locking exists (see below for further analysis).

On the other hand, the occurrence of a magnitude 8.3–8.5 earthquake means that strain must accumulate somewhere. There is no known place where a fault that slips predominantly aseismically also fails in occasional great earthquakes, but a natural way to reconcile the accumulation of elastic strain with aseismic slip on the plate interface would be for that slip rate to be varying with depth. Variation in slip rate would cause strain to accumulate in the surrounding volume, which would eventually be released in an earthquake. One such variation is illustrated in Fig. 2.12, which shows that an increase in slip rate on the plate interface over the depth interval 45–70 km would cause compressional strain to accumulate beneath Crete.

Variation of slip rate with depth might result from a contrast in mechanical properties of different parts of the overriding plate, for instance between the accretionary prism of sediments and the continental crust of the Aegean. The combination of faulting suggested here may be common in subduction zone settings. A thrust fault has been imaged in the Nankai trench off Japan at the landward edge of the accretionary prism [30, 31] and slip on a similar fault is thought to have caused the 10 m uplift of Middleton Island during the 1964 Alaska earthquake [32, 33].

Repeated slip of the kind I propose for the AD 365 earthquake would cause permanent uplift of the surface of Crete by thickening the pile of sediments beneath it [34], and may account for the low seismic wave speeds beneath the island [35]. The contrast between the steep southern coast of Crete and its gentle northern slope (Fig. 2.7, Chap. 5) is consistent with tilting by uplift on a fault cropping out near the south coast, so this proposed fault may control both the uplift and the extent of western Crete itself. Indeed, this geometry of faulting may be a common cause of long-term coastal uplift above subduction zones.

In summary, I conclude that the AD 365 earthquake took place by 20 m of slip on a fault ~100 km long, dipping NE at 30° from the Hellenic Trench to a depth of 45 km. It has long been thought that slip along the plate interface between the Aegean and the Mediterranean ocean floor is predominantly aseismic. My arguments here support that view but emphasise that the Hellenic Trench, which is a major feature along 400 km of the western coast of Greece, appears to mark the outcrop of a separate fault that is capable of producing rare, very large, tsunamigenic earthquakes (see also [8]). In the following section, I attempt to assess the likelihood of a repetition of an event like the AD 365 earthquake.

2.7 Future Tsunamigenic Earthquakes in the Hellenic Subduction Zone

It is possible to estimate the amount of elastic strain accumulating near the Hellenic trench by measuring relative motion across the Southern Aegean using a continuous GPS network. The GPS stations whose data are used here are part of a network of

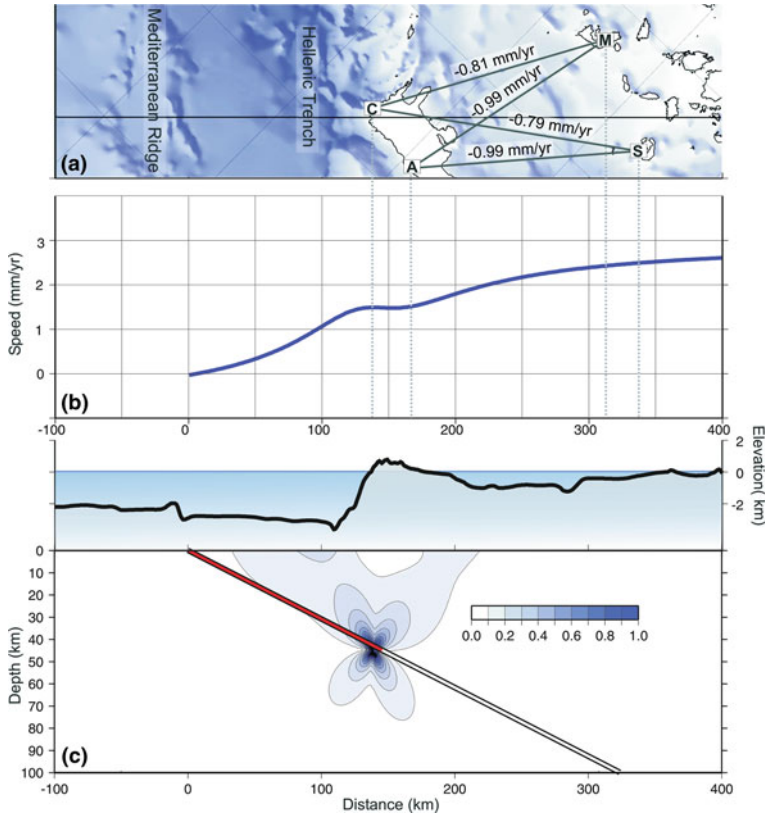


Fig. 2.12 **a** Continuous GPS data show shortening of ~ 1 mm/year between Chrisoskalitissa (C) and Anopoli (A) in SW Crete and Milos (M) and Santorini (S), ~ 200 km to the NE. **b** South-westward speed, calculated from model illustrated in (c), of points within the Aegean, relative to the Mediterranean Ridge. **c** Distribution of shear stresses calculated [17] for a planar fault, representing the interface between the African plate and the Aegean lithosphere, that slips 4 mm/year more rapidly below 45 km (*open line*) than above that depth (*red line*). Relative values of shear stress are shown by the *colour bar*; absolute values, at any particular time, depend linearly on the slip difference that has accumulated between the two fault segments

20 continuous stations run by the National Technical University of Athens and the University of Oxford. Data was processed by M. Floyd. Lines joining stations in western Crete to Milos and Santorini, about 200 km to the north, are shortening by approximately 0.9 mm/year (Fig. 2.12a), with an uncertainty of about 0.1 mm/year. Such shortening is a small fraction of the total Nubia-Aegean convergence, confirming that the majority of the convergence is taken up by steady sliding along the subduction zone interface. However, this small amount of contraction can be explained by locking on a fault to the south-west of Crete. If we assume that elastic strain builds up because of a variation of slip rate with depth, we can calculate what proportion of the 35 mm/year convergence is going into locking. The observed

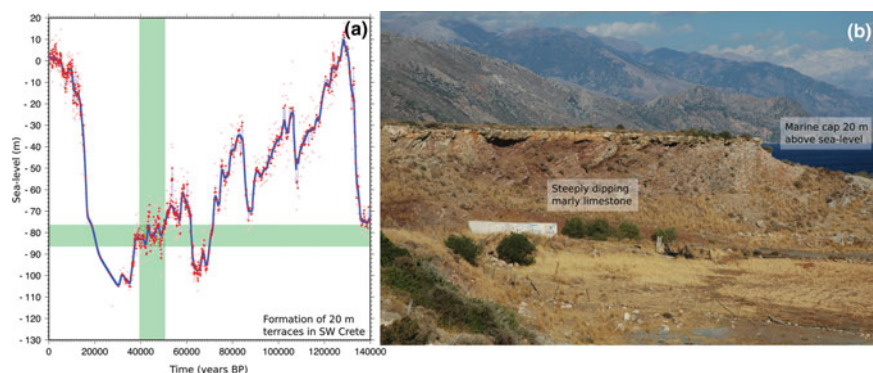


Fig. 2.13 **a** Sea-level curve plotted for the last 140,000 years from Rohling et al. [37]. Small *pink dots* are raw data points, and the *red dots* result from their application of a three-point moving average. The *pale blue line* is my interpolation between these *red points*, using an Akima spline [44]. The *dark blue line* shows the result of applying a Gaussian filter of width 3,000 years to this interpolated curve. The radiocarbon dates of bivalves collected from terraces at 20 m elevation from Paleochora and Chrisoskalitissa suggest that these terraces formed between 40,000 and 50,000 years BP. At this time, sea-level was 75–85 m below present day sea-level (*green band*). **b** Photograph of the 20 m terrace at Paleochora (35° 13.626'N, 23° 40.895'E). The tilted country rock is unconformably overlain by a horizontal marine cap from which the bivalve sample was taken

contraction is consistent with a slip deficit of ~ 4 mm/yr on the upper part of the subduction interface (Fig. 2.12b); in other words about 10% of the total convergence rate is being stored as elastic strain, with the rest being taken up by aseismic slip. At this rate it would take $\sim 5,000$ years to accumulate strain equivalent to the 20 m of slip that was released in the AD 365 earthquake. The predecessor of the AD 365 earthquake would then, plausibly, have happened more than 6,000 years ago, when sea-level was tens of metres below its present position [36, 37].

Any direct evidence of uplift in an earthquake a few thousand years before 6,000 years BP would now be submerged, but some support for this analysis comes from marine terraces at 20–22 m present elevation in western Crete (see Sect. 5.5). Bivalves were collected from the marine encrustation at the top of the terraces, and were radiocarbon dated. These AMS ^{14}C dates show the terraces were formed 40–50,000 years BP (Table 2.1, Fig. 2.13), when the sea surface was ~ 75 –85 m below its present level [36, 37]. Therefore ~ 100 m uplift has occurred in 40,000–50,000 years. This observation independently suggests a long-term average uplift rate of ~ 2.0 – 2.5 mm/yr, which is consistent with uplift occurring in AD 365-type events that repeat every $\sim 4,500$ years (4,000–5,000 years). See Chap. 5 for further discussion.

The processes described here for western Crete may also occur along the rest of the Hellenic subduction zone. The AD 1303 earthquake and tsunami are thought to have originated near Rhodes [1, 38], so the whole Hellenic subduction zone may represent a tsunami hazard for the eastern Mediterranean. If the partitioning between aseismic and seismic slip is the same along the whole zone as in the 100 km section

near SW Crete, an AD 365-type earthquake should be expected every ~ 800 years. That there has been only one other such known event (in AD 1303) in the past 1,650 years should focus our attention on the modern-day tsunami hazard in the Eastern Mediterranean.

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Active tectonics of the Hellenic subduction zone

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2012, XIII, 169 p., Hardcover

ISBN: 978-3-642-20803-4