Field Trip Guide, 2011

Editors

Active Tectonics and Earthquake Geology of the Perachora Peninsula and the Area of the Isthmus, Corinth Gulf, Greece

G. Roberts, I. Papanikolaou, A. Vött, D. Pantosti and H. Hadler

2nd INQUA-IGCP 567 International Workshop on Active Tectonics, Earthquake Geology, Archaeology and Engineering

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Preface

After the very successful 1st Workshop on Earthquake Archaeology and Paleoseismology held in the ancient roman site of Baelo Claudia (Spain, 2009), the INQUA Focus Group on Paleoseismology and Active Tectonics decided to elaborate a bi-annual calendar to support this joint initiative with the IGCP-567 "Earthquake Archaeology". This second joint meeting moved to the eastern Mediterranean, a tectonically active setting within the Africa-Eurasia collision zone and located in the origins of the pioneer's works on archaeoseismology. However, for the coming year 2012, at least a part of us will move also to the New World, where the 3rd INQUA-IGCP 567 international workshop will take place in Morelia, Mexico in November 2012. It is planned to proceed with the meeting, so we are thinking of Aachen, Germany, to be the host in 2013, possibly together with Louvain, Belgium.

The aim of this joint meeting is to stimulate the already emerging comparative discussion among Earthquake Environmental Effects (EEE) and Earthquake Archaeoseismological Effects (EAE) in order to elaborate comprehensive classifications for future cataloguing and parametrization of ancient earthquakes and palaeoearthquakes. One of the final goals our collaborative workshops is the integration of archaeoseismological data in Macroseismic Scales such as the Environmental Seismic Intensity Scale ESI-2007 developed within the frame of the International Union for Quaternary Research (INQUA). In this second workshop we offer again a multidisciplinary and cross-disciplinary approach and program, since there is an urgent necessity to share the knowledge and objectives among geologists, seismologists, geodesists, archaeologists and civil engineers in order to improve seismic hazard assessments and analyses in a near future. Also, we intend to sharpen geoscientists and their research more in the direction of critical facilities, which are of world-wide public and political interest after the dramatic catastroph in Fukushima, Japan.

The last two years provided significant dreadful earthquake scenarios, which were in most of the cases oversized in relation to the data provided by the historical and instrumental seismicity. The Haiti Mw 7.0 (Haiti, Jan 2010), Malua Mw 8.8 (Chile, May 2010), Christchurch Mw 6.3 (New Zealand, Feb 2011), Tohoku Mw 9.0 (Japan, Mar 2011) and Lorca Mw 5.1 (Spain, May 2011) events illustrates that both extreme subduction earthquakes or moderate events can generate severe damage in relation to relevant secondary coseismic effects or Earthquake Environmental Effects (EEE). Most of these recent events have clearly demonstrated that the vibratory ground shaking is not the unique, or even most significant, source of direct damage, and it is by no means the only parameter that should be considered in seismic hazard assessments. The lessons offered by the aforementioned events corroborate once again the relevance of liquefaction, tsunamis, rockfalls, landslides, ground subsidence, uplift or failure as a major source of hazard. But this also underpins the need of re-evaluating the significance of macroseismic intensity as an empirical measurement of earthquake size. In fact, as highlighted in the last volume produced by the INQUA Focus Area (Serva et al., 2011), intensity is a parameter able to describe a complete earthquake scenario, based on direct field observation and suitable to be preserved in the geological, geomorphological and archaeological records.

With this aim the INQUA TERPRO #0418 Project (2008-2011) has implemented a world-wide online EEE Catalogue based on Google Earth in order to promote the use of the ESI-2007 Scale for seismic hazard purposes www.eeecatalog.sinanet.apat.it/terremoti/index.php. On the other hand the IGCP-567 is promoting an interesting shared approach of EEE data and EAE data for the same purpose. Examples of this variety of original research coming from this collaborative approach are the Geological Society of London Special Volume 316 (2009) *Paleoseismology: Historical and Prehistorical records of Earthquake Ground Effects for Seismic Hazard Assessment* (K. Reicherter, A.M. Michetti & P.G. Silva, Eds.), the Geological Society of America Special Papers 471, *Ancient Earthquakes* (2010) (M. Sintubin, I.S. Stewart, T. Niemi & E. Altunel, Eds.) and the Special Volume of Quaternary International (2011) *Earthquake Archaeology and Paleoseismology* (P.G. Silva, M. Sintubin & K. Reicherter, Eds.). In the same way, this abstract volume contains more than 80 contributions from researchers of more than 27 different countries and illustrates the upgrading shared knowledge on palaeo-, ancient, historical and instrumental earthquakes and images an impressive growth of our community. Our workshop was co-ordinated through the newly established website www.paleoseismicity.org, where earthquake info and blogs are openly shared.

Finally, we wish all participants a fruitful conference and workshop in the vicinity of the ancient sites of the Classical Greece around the Corinth Gulf, where earthquake science, wonderful landscapes, ancient cultures, amazing sunny days, fantastic "Greek cooking", nice beaches, daily cool beers, wine tasting events and late night gin tonics mixed with hot discussions are waiting for all of us. A special "**efharisto poli**" goes to Christoph Grützner and Raul Pérez-López for their invaluable work with the organisation and Georgios Deligiannakis for the handling of this fieldtrip guide.

The Organizers of the 2nd INQUA-IGCP 567 Workshop

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Field Trip Day 1

Fault evolution in the Gulf of Corinth implied by uplifted palaeoshorelines and faults scarps.

by Gerald P. Roberts¹ and Ioannis D. Papanikolaou^{1,2}

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Abstract

The Corinth Rift has narrowed through time, with faults north and south of the gulf progressively abandoned (Jackson 1999) (Fig. 1). Extension is now concentrated on the southern shores of the Gulf of Corinth, but was formerly (Pliocene and early in the Quaternary) widespread throughout central Greece (Roberts et al. 2009). We will examine outcrops that may explain why faulting localised (Fig.2). This field trip will examine locations on the Perachora Peninsula which forms the western tip of the Pisia-Skinos Fault. We will examine fault traces, dated carbonate and clastic sediments and wind-gaps (palaeo-valleys) that record the westward propagation of the tip of the Pisia-Skinos Fault. The timing of events gained from U-Series coral dates will allow us to examine how, when and possibly why strain has localised into ~30 km wide Gulf of Corinth when the rift was formerly many times wider in the Quaternary. The reason why localisation occurred may be explained by alongstrike fault propagation (Morewood and Roberts 1999, Roberts et al. 2009). For example, when the surface expression of the Pisia-Skinos fault achieved its present length, footwall uplift rates increased by a factor of 3.2 ±0.2 at 175 ± 75 ka, synchronous with the death of the across-strike Derveni fault at 382-112 ka (Flotte et al. 2001). Before 175 ± 75 ka, both the Pisia-Skinos Fault and the Derveni Fault were active. After 175 ±75 ka, death of the Derveni fault and other faults may have transferred extra regional strain per unit time onto the Pisia-Skinos fault (and perhaps other faults), forcing it to accelerate and change the spatial distribution of strain-rates responsible for the velocity field. This style of rift evolution where one set of localised faults developed within a more distributed fault network with a resultant increase in slip-rate on faults that remain active appears to typify areas of extension, such as the North Sea (Cowie et al. 2005).

Geological background

Active extension of continental crust in central and southern Greece occurs in the zone of convergence between the African and Eurasian Plates (Fig. 1). Classically, this has been viewed as resulting from trench roll-back of the Hellenic subduction zone which is consuming the last remaining portions of Tethyan ocean crust (back-arc extension). The area is also affected by deformation at the tip of the propagating North Anatolian Fault. Extension occurs in an approximately N-S direction in the area containing the c. WNW-ESE striking marine basins of the Gulfs of Corinth and Evia. Extension occurs in an area previously thickened as an alpine fold-thrust belt, with thrusts striking NNW-SSE. Thus, normal faults are almost at right-angles to pre-existing thrusts. This area lies at the western end of the right-lateral strike-slip North Anatolian Fault. Present-day extension rates of ~10 mm/yr occur across the Gulf of Corinth, with rates that are probably lower (~2-3 mm/yr) across the Gulf of Evia. The age of extension in the Gulfs of Evia and Corinth are controversial, but certainly include Plio-Pleistocene syn-rift sediments. Pre-rift rocks include Miocene-Oligocene clastic sediments, Triassic to Oligocene carbonates with ophiolitic thrust sheets, all underlain by metamorphic schists. Active volcanic activity occurs in the nearby Saronic Gulf with the active volcano, Methana. Surface ruptures with c. 1 m vertical offsets occurred along basin-bounding normal faults during the two 1894 Evia earthquakes (>Ms 6.0), and the three 1981 Gulf of Corinth earthquakes (Ms 6.7, 6.4, 6.3). Earthquakes have produced dramatic coastal uplift and subsidence. The most striking geological aspects of the Gulf of Corinth (see Armijo et al. 1996, Jackson 1994) are:

1) the extension is very rapid with individual faults having slip-rates as high as c. 10 mm/yr);

2) there have been 11 earthquake with magnitudes > Ms 6.0 in the last c. 100 years; - the entire extensional basin formed in as little as 2-4 Myrs; within this time period the faulting has localised from a zone of 2-3 parallel active faults on to a single active fault in the central gulf;

3) the basin is underfilled with sediment and has c. 800 m water depth at its centre;

4) the southern margin shows dramatic uplift due to presently-active offshore faults, with Plio-Pleistocene marine sediments at elevations of upto 1200 m;

5) Quaternary sea-level changes of c. 120 m every c. 100 kyrs, together with water depths of only 60 metres at the western end of the gulf (the Rio Sill), mean the gulf turns into a lake at glacio-eustatic sea-level lowstands.

The GPS velocity field in the Gulf of Corinth (e.g. Briole et al. 2000) shows extension is localised on northdipping active normal faults on the south side of the gulf. Seismic reflection data and observations onland show that other normal faults that were active probably in the Pliocene have ceased activity.



When exactly did faults change activity-rates and how did this change the velocity field? Can we date changes in the velocity field? What does this imply for the age of the faulting?





Fig. 2: Location map for localities to be visited.

Stop 1

The 1981 Earthquake sequence -Megara Basin – Alepochori village

Ioannis Papanikolaou and Gerald Roberts

The 1981 Alkyonides Earthquake sequence

On February 24, 25 and March 4, 1981 three (Ms=6.7, Ms=6.4, Ms=6.3) successive destructive events (20 fatalities and 500 injured) occurred at the eastern end of the Corinth Gulf (Fig.3) (Jackson et al. 1982; Papazachos et al. 1982; Taymaz et al. 1991). The last two events of the sequence lie in areas where a positive Coulomb stress increase has been calculated (Hubert et al. (1996)), implying that this earthquake sequence was the result of stress transfer that triggered the second and third events. All three events correspond to normal faulting, accommodating N-S extension. The focal mechanisms that described the co-seismic slip at depth (~10 km), exhibit similar fault plane orientations and kinematics to those measured on the faults at the surface (Morewood and Roberts 2001).



Fig. 3: The 1981 Alkyonides earthquake sequence eastern Corinth Gulf. View of the epicentral region with emphasis on the primary surface ruptures and coastal uplift/subsidence. Sketch modified from Jackson et al. (1982), Mariolakos et al. (1982) and Hubert et al. (1996).

Damages were widespread in three different provinces (Beotia, Attica and Corinth) where in total 7,701 buildings collapsed or had damages beyond repair and 20,954 buildings were severely damaged (Antonaki et al. 1988). These events produced numerous earthquake environmental effects (EEE) such as rockfalls, landslides (both onshore and offshore), liquefaction, a weak tsunami wave, significant coastal subsidence and uplift, but most importantly extensive primary surface faulting. In particular, in the Pisia Fault, surface ruptures were longer than 10 km and displacements were in the range of 50-70 cm with a maximum recorded value of 150 cm, whereas in the Skinos (or Schinos)

fault segment displacement was about 100 cm (Jackson et al. 1982; Fig.3). The March 4 event ruptured the Plataie-Kaparelli Fault Zone (~10 km of surface ruptures) producing an average 50-60 cm of throw (Jackson et al. 1982; Mariolakos et al. 1982, Fig.4) and a maximum heave and throw of 60 and 120 cm respectively between Kaparelli and Plataies (Papazachos et al. 1982).



Fig. 4: West of Alepochori up to the western part of the bay of Strava 60 cm of subsidence was observed, flooding up to 50 m of the former shore (Mariolakos et al. 1982). b) View of the surface ruptures on the Plataies-Kaparelli fault zone during the March 4 event, producing 50-60 cm of throw (70 cm of displacement).

West of Alepochori up to the western part of the Bay of Strava, 60 cm of subsidence was observed, flooding up to 50 m of the former shore (Mariolakos et al. 1982, Fig.3, Fig.4), whereas, east of Alepochori coastal uplift was also observed (Jackson et al. 1982; Vita-Finzi and King, 1985). In Schinos and Strava coastlines, there was a disagreement on the amount of subsidence recorded ranging from 50-80 cm (Andronopoulos et al. 1982; Mariolakos et al. 1982; Hubert et al. 1996) up to 120 and 150 cm (Khoury et al. 1983; Vita-Finzi and King, 1985), but based on Hubert et al. (1996) arguments and modelling, values higher than 100-120 cm probably overestimate the co-seismic effect. Submarine slumping in the Alkyonides deep basin and several mass-movement phenomena in the shelf area have also been detected such as a large-scale slump about 10 km long, 1.5-2 km wide, extending 16 km² over a depth of 360 m (Perissoratis et al. 1984). Jackson et al. (1982) quoted that local people reported a 1 m high tsunami during the main shock in the Alkyonides Gulf, therefore, it is possible that the tsunami generation can be attributed to the large scale slumping detected by Perissoratis et al. (1984).



Fig. 5: MS (Mercalli-Sieberg, a version similar to the MCS) intensity distribution of the 1981 Alkyonides earthquake sequence (BGINOA 1981, Antonaki et al 1988).

These shallow normal faulting earthquakes left their mark on the country and the population since they affected not only the Perachora Peninsula (Maximum intensity IX-X), Plataies (IX-X) or Kaparelli (IX), but also the city of Athens that is located 70 km to the East, where tens of buildings collapsed in certain town districts (Fig.5). This earthquake sequence had a core of high intensities around the epicentral area and a second maximum of high intensities at 70 km distance, affecting several districts of Athens. On average, Athens experienced intensities VII and VIII, however, in some boroughs and building blocks, intensities up to IX were also recorded. In particular, in Athens 1,175 buildings collapsed or had to be demolished after the earthquake, whereas 7,824 buildings experienced severe damage (Antonaki et al. 1988). Following the 1981 earthquake sequence an updated and stricter building code was implemented across the country.

The 1981 earthquake sequence and the ESI 2007

All earthquake environmental effects have been assessed according to the Environmental Seismic Intensity scale ESI 2007 (Michetti et al. 2007). A maximum epicentral ESI 2007 value of X is determined in several sites (Fig.6) and particularly along the strike of the primary surface ruptures in Pisia, Skinos and in Kaparelli - Plataies where they exceeded a few tens of cm in height. Intensity X has also been allocated along the coastal zone from Strava up to Alepochori, where significant subsidence ranging from a few decimetres up to 100 cm have been recorded. Intensity IX is mainly assigned in areas where the surface ruptures were a few tens of centimetres high. Finally, intensity VIII was widely assigned affecting a large area where ground ruptures, extensive landslides, rockfalls and liquefaction phenomena have been observed.



Fig. 6: View of the Mercalli-Sieberg (Antonaki et al. 1988) and the ESI 2007 (Papanikolaou et al. 2009) intensity scales in the epicetral area.



Fig. 7: Comparison of the isoseismal pattern between the MS and the ESI 2007 intensity scales (Papanikolaou et al. 2009).

Maximum MS (Mercalli-Sieberg) intensity values (a version similar to the MCS) were recorded in all villages that were in close proximity to the activated faults (Fig.6; Perachora IX-X, Plataies IX-X, Skinos IX, Pisia IX, Kaparelli IX). However, no intensity X has been assigned and most of the epicentral villages recorded an epicentral intensity IX. Figure 7 shows the different isoseismal pattern of the epicentral region based on the ESI 2007 and the Mercalli-Sieberg scales, respectively. It should be noted that surface geology played a decisive role in the damage distribution and had a significant effect on the intensity observed at a site. On average, a survey in the area showed (Tilford et al. 1985) that under similar circumstances sites located on soil foundations experienced about one intensity degree more shaking than sites located on rock foundations, whereas sites on Neogene sediments experienced about half degree greater intensity than sites located on rock foundations.

An important issue regarding the 1981 Alkyonides earthquake sequence is the different isoseismal distributions recorded by several research groups. In several epicentral villages reported intensity values that differ from half (e.g. Perachora and Pisia) up to one degree (Skinos). For example, in Skinos village, intensity recordings varied from VIII-IX (Papazachos et al. 1982), IX (Bulletin of the Geodynamic Institute of the National Observatory of Athens - BGINOA 1981) up to IX-X (Andronopoulos et al. 1982), providing a rather confusing pattern. This difference can be attributed to several causes. It may be: i) due to the subjective interpretation of damages or ii) due to the subjectivity in allocating the predominant damage in a site, or iii) because the assigned intensities correspond to the maximum observed intensity rather than the mean. From this perspective, the ESI 2007 scale is probably easier to implement, more precise in quantifying macroseismic effects, offering a higher objectivity in the process of assigning intensity values.

In the 1981 Alkyonides example, the ESI 2007 intensity scale provides not only a slightly higher maximum epicentral intensity (X), but also a different spatial distribution of the isoseismals, compared to the traditional scales (Papanikolaou et al. 2009). This implies that current traditional scales possibly underestimate the "strength" of this kind of earthquake sequence (e.g. Serva 1994). This partly occurs because the epicentral area, where significant EEE were recorded, was sparsely populated. In addition, several villages located in the epicentral region were founded on bedrock sites and others despite been proximal to the surface ruptures were located on the footwall block, such as the Kaparelli or Alepochori villages, experiencing less shaking and damages. The 1981 earthquake sequence emphasizes the importance and the increasing accuracy of the ESI 2007 scale towards the highest levels of the scale in the epicentral area.

The Megara Basin

Megara is a Neogene basin with a lower sequence of Upper Miocene and an upper sequence of Pliocene and Quaternary marine and lacustrine sediments (Theodoropoulos 1968). It forms a tectonic semi-graben in between the horsts of Gerania (1.351m) to the southeast and Pateras (1.432m) to the northwest. Strata within the basin are titled 10-40° to the NNE so that a WNW-ESE rotating axis is deduced (Fig. 8, Mariolakos and Papanikolaou 1982). This rotation relates to the WNW-ESE trending normal fault that bounds northeastwards the Megara Basin. This fault was active during Pliocene and Lower Pleistocene times, but at present day shows no sign of activity and has been offset by the ENE-WSW Skinos-Psatha fault segments that were partly ruptured during the 1981 earthquake sequence (Mariolakos and Papanikolaou 1982, Leeder et al. 1991). The Psatha segment lies at the eastern termination of the South Alkyonides fault system where finite throws, footwall uplift and hangingwall subsidence are at a minimum and progressively increase towards Perachora (Roberts and Gawthorpe 1995). However, no surface faulting was recorded onshore at

Psatha in 1981 (Jackson et al. 1982). In conclusion, the WNW-ESE fault that bounds Pateras Mt is now inactive and predates the activity on the South Alkyonides fault system. The ENE-WSW trending Skinos and Psatha segments of the South Alkyonides fault system uplift the eastern boundary of the Megara basin as evidenced by the 1981 earthquakes, implying that another younger titling is evidenced around a ENE-WSW axis rotation.



Fig. 8: Schematic block diagram illustrating the rotations of the tectonic blocks and the resulting morphological features (modified from Mariolakos and Papanikolaou 1982).

The drainage divide within the Megara basin is not located near the centre of the basin, but is shifted to the east due to the ongoing uplift of the Psatha - Skinos fault segments. Therefore the drainage network that flows towards the Saronic Gulf is the dominant system, draining about 78% of the basin in comparison to the 22% that flows to the Corinth Gulf (Fig.9).



Fig. 9: View of the drainage network in the Megara basin. 78% of the basin is drained towards the Saronikos Gulf. The drainage divide is shifted towards the Corinth Gulf due to the ongoing uplift of the Psatha - Skinos fault segments.

The Megara basin progressively became inactive and abandoned with initiation of the younger South Alkyonides fault system and the development of the Alkyonides basin (Mariolakos and Papanikolaou 1982, Leeder et al. 1991). This is also confirmed by paleomagnetic studies distinguishing differential rotational domains which show that the Megara region belonged to the Beotia-Locris block in the past, but has now been incorporated into the Peloponnesus block, because the faulting in the Gulf of Corinth has propagated both north and east. This is revealed by the discrepancy between GPS-inferred rotations and paleomagnetic rotations on a scale of one million years (Mattei et al. 2004). Sakellariou et al. (2007) based on offshore sequence stratigraphy suggest that the onset of the Alkyonides basin subsidence initiated at 0.4-0.45 Myrs ago. However, there is a debate regarding the inferred onset of fault propagation in the eastern Corinth rift and the Megara basin fault abandonment with timing ranging from less than 1Myrs, to 1Myrs (Colllier et al. 1992, Armijo et al. 1996) or lately into 2.2 Myrs (Leeder et al. 2008).

The Alepochori Village

The Alepochori village lies in the immediate footwall of the Skinos fault and was coseismically uplifted approximately 20cm during the 1981 events. This can explain why it experienced less damages (intensity VIII) compared to other villages in the epicentral area. The fault can traced offshore a several hundrend meters westewards of Alepochori (Sakellariou et al. 2007) and links to the Psatha segment. This ongoing uplift is also evident from the relict rocks and notches traced along the coastline from Alepochori village to the Bambakies-Skinos fan (Fig. 10).



Fig. 10: View the notches in an uplifted limestone block along the Alepochori –Skinos coastal road.

Stop 2

Paleoseismological trenching at the Bambakies Fan: in search of 1981 earthquake ancestors

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1.Introduction

In 1981, a series of three earthquakes (February 24, 1981, 6.7 M_S ; February 25, 1981, 6.4 M_S ; March 4, 1981, 6.4 M_S) struck the Alkyonides Basin in the eastern Gulf of Corinth [*Jackson et al.*, 1982; *King et al.*, 1985; *Abercrombie et al.*, 1995]. Surface faulting occurred on the Pisia and Skinos Faults on the Perachora peninsula and on the antithetic Kaparelli Fault in the northeastern part of the basin (Fig. 11). Whereas the antithetic Kaparelli surface rupture was unambiguously produced by the third shock, there remains uncertainty as to whether the Pisia and Skinos ruptures should be ascribed to one or both of the first two earthquakes, as both occurred during the same night.

The Pisia and Skinos Faults form the onshore portion of the echelon set of normal faults which comprise the main Alkyonides Gulf basin-bounding fault system about 35 km long. (Fig. 11). It is unknown whether there was any surface break associated with the 1981 earthquakes on the submarine fault scarps. Surface faulting on the Pisia and Skinos Faults occurred along preexisting scarps in crystalline carbonates, dunites, and adjacent alluvium and colluvium. Displacements of up to 1.4 m were measured by *Jackson et al.* [1982]. Hanging wall subsidence during the 1981 earthquakes was expressed by inundation of the coastline around the villages of Skinos and Psatha [*Jackson et al.*, 1982].



Fig. 11: Location of the Bambakies Fan in the eastern end of the Corinth Gulf (inset). The three 1981 mainshocks [Jackson et al., 1982] and the trace of the main active faults that have controlled the subsidence of the Alkyonides Gulf are shown, red ones are those that ruptured the surface in 1981.

The coincidence of the 1981 deformation with topography confirms that the 1981 events represent the dominant tectonic process in the area producing important subsidence in the fault hanghingwall testified by water depths of up to 360 m. Footwall uplift is evidenced by raised Holocene beachrock, raised abrasional and biogenic notches of Holocene age [*Pirazzoli et al.*, 1994; *Stewart and Vita-Finzi*, 1996], raised late Pleistocene marine terraces, and the occurrence of marine-influenced sediments within deposits of the older Megara rift basin, now raised above sea level [*Leeder et al.*, 1991; *Collier et al.*, 1992].

The evidence of their long-term activity makes the Skinos and Pisia faults good potential targets for paleoseismological research. However, the limited amount of locations with recent (Holocene and Upper Pleistocene) deposits at both sides of the 1981 ruptures made the paleoseismological site selection quite difficult. Because of the recentness of its deposits and the presence of compound fault scarps, indicating repeated movements on the same fault section, we opened three trenches on the Bambakies fan, at the eastern end of the Skinos Fault 1981 surface breaks (Figs 11 and 12). The Bambakies fan consists mainly of pebble to boulder of ultrabasic, dunite composition, derived from the erosion of the fault footwall, with limited coarse matrix. The upper oldest surface of the fan was likely attained around 6,000 yr ago, when sea level approached the present elevation. Now this surface is entrenched by the active channel and indented by several lower, younger surfaces. Nowadays, the main fluvial channel runs north to northwest down the Bambakies Fan and is active only following storms. The paleoseismological study of the Bambakies fan was performed in 1996 and results are published in two papers where it is possible to retreive more details: Pantosti et al. (1996) and Collier et al (1998).



Fig. 12: 1981 surface ruptures at the Bambakies Fan. Up to three en echelon surface breaks can be traced from a basement-bounding fault contact to the west of the Bambakies fan, in a linear deformation zone approximately 30-40 m wide, typically 110° striking. The fault cannot be traced across the basement ridge to the east of the Bambakies Fan; displacement is probably transferred to offshore faults to the north. 1,2, and 3 indicate location of the paleoseismological trenches; Sites A and B are the topgraphic survey areas (see Figs 13 and 14); a to e are topographic profiles (see Figs 13 and 14).

2. Microtopography

We carried out a detailed microtopographic survey to characterize the fault scarp across the fan surface. Scarp morphology clearly shows that the total scarp height is increasing with the age of the faulted fan surface (from 0.5 m up to ~6 m - Figs 13 and 14). In most of the cases, 1981 ruptures run at the base of the cumulative scarp (Photo 1) and show average throws of 0.7 m, with local maxima of 1.3 m. No evidence for seizable horizontal movement was found on the basis of the mapping of piercing points.



Fig. 13: Topographic map of Site A (for location see figure 2) highlighting different fan surfaces displaced by the fault with the older ones recording higher displacements resulting from subsequent surface faulting events.



Photo 1: View of a cumulative scarp near site B comprising the 1981 free-face and an older retreated scarp.

At the eastern part of the fan, the main scarp is faced, at a distance of about 30 m, by a 1 m high antithetic scarp (Fig. 12). The 1981 ruptures appeared as a ~0.5 vertical break at the main scarp and as a decimetric-high rupture at the antithetic. This implies the occurrence of several prior earthquakes displacing the same fan surface. Assuming an age of 6,000 yr for the displaced fan surface and taking the 6 m high scarp we can obtain a reference vertical separation rate of 1 mm/yr.



Fig. 14: Profiles across differently displaced fan surfaces at or near site A highlighting also the presence of the antithetic scarp; trace of profiles in Figs 12 and 13.

3.The trenches

3.1 Trench 1

This trench (Photo 2) lies west of the modern Bambakies Fan channel (Fig. 12) and exposes the fault zone which generated the main 1981 earthquake scarp at this location and a secondary fault zone 5 m to the north, which displayed only minor warping and displacement in 1981 (Fig. 15). The trench cuts obliquely across the scarps that, at this particular location, are almost parallel to the alluvial channel flow. This produces a complex interaction between alluvial and tectonic-related geomorphology. The deposits exposed in the trench wall comprise poorly sorted, pebble, and cobble conglomerates, with silty or sandy matrix, and open framework deposits (alluvial units: 3, 5, and 7) along with two important paleosols (units 4 and 6). The offset of these paleosols and stratigraphic correlations across the two fault zones are at the basis of the recognition of past faulting events (eh in Fig 16).



Photo 2: View of trench 1 (before the deepening of the secondary fault zone (left).

The 1981 earthquake (event 1) is represented by a total vertical displacement across the two fault zones of 0.5-0.7 m. Post-1981 deposits are limited to fissure-fill material immediately below the free face. Event 2 is recognized after removing the 1981 deformation and is characterized by a warping coincident with the main fault scarp and a vertical offset of ~0.5 m of paleosol 4 across the secondary fault zone. The preferred event 2 horizon (i.e., the topographic surface at the time of the earthquake) sits directly upon paleosol 4, which clearly predates the event, but stratigraphic relationships across the secondary fault scarp are not definitive and would allow the event horizon to overlie the lower deposits of alluvial unit 3.



Fig. 15: Log of trench 1 from a 1:20 scale survey.Dotted lines labeled eh are event horizons, rectengles enclose dated samples and are connected to their cal. age. Units 1, 4 and 6a are organic soil layers, the remaining units are alluvial fan deposits.



Fig. 16: Sequential reconstruction of the sediments and structures exposed in the trench as they would have appeared prior to the 1981 earthquake and prior to past events. Note that uncertainties in estimation of past displacements arise because of uncertainties in the original form of the fan surface and because of the fault intercepting the surface as multiple scarps at this location, such that some paleoearthquakes may be missed by Trench 1.

Event 3 is revealed by removal of event 2 displacement too. In this way paleosol 4 is seen to have developed on a surface with a marked channel morphology and beneath this, unit 5c and paleosol 6 show a ~1.2 m net vertical offset produced by this event (~0.7 m on the secondary fault zone and ~0.5 m on the main).

Summarizing, three surface faulting events are evident in Trench 1, including the 1981 earthquake. Each event exhibits a local vertical throw in the range 0.5-1.2 m. The age of these events was set by means of AMS and conventional radiocarbon dating of charcoal fragments and paleosols. Event 2 is older than 1650-1825 A.D. (sample TR1-C5) and younger than 1295-1495 A.D. (sample TR1-S3). This suggests that event 2 occurred in the interval 1295-1825 A.D., with the earlier part of this range the more likely, taking our preferred stratigraphic position of event horizon 2 with respect to the dated samples. Event 3 is constrained by 390 A.D.-895 A.D. (conservatively derived from 590-895 A.D. - sample TR1-S2 and 390-630 A.D. - TR1-20) and 800-1015 A.D. (sample NTR1-S2). Event 3 therefore occurred in the interval 390-1015 A.D. and possibly 590-1015 A.D.

3.2 Trench 2

Trench 2 is located 40 m east of the active channel on the Bambakies Fan, where the fault has displaced the fan surface by 3.5 - 5.0 m (Figs 12, 13, 17). Here the 1981 free face (Photos 3, 4) is up to 0.6 m high. Excavation revealed a sequence of alluvial fan deposits buried by scarp-derived colluvium and interbedded paleosols. A single fault zone crops out in the east wall of the trench, whereas a small graben is seen in the west wall, where the main fault is faced by a steeply dipping antithetic normal fault 1.5 m to the north (Fig. 17).



Photo 3: View of the 1981 free-face cutting the youngest alluvial surface, just east of the active channel.

Photo 4: Detail of the fault zone of trench 2. Notice the colluvial wedges piling up in the hanging-wall.

Evidence for several individual surface faulting events can be found on the basis of individual colluvial wedges and stepwise increases of displacement with depth across the antithetic fault. Uncertainty arises from alternative interpretations of colluvial wedges having been produced either by collapse and erosion of a newly generated scarp or by sporadic storm erosion of a preexisting scarp. The absence of stratigraphic markers in the footwall correlating with those in the hanging wall prevents us from determining throws of individual faulting events prior to 1981. However, on the basis of the correlation of unit 9 with the soil still forming on top of the footwall, we can ascribe the 3.5 - 5.0 m of cumulative vertical throw (profile c, Fig. 14) to 6-7 discernible events. This gives an average vertical throw of 0.5-0.8 m per surface-breaking event.



Fig. 17: Log of trench 2 from a 1:20 scale survey.Colored lines highlight possible event horizons set on the basis of the occurrence of colluvial wedges and increase of displacement with depth on the antithetic fault zone.

Because of the lack of datable material other than scarp-derived, likely re-worked organic matter enclosed in the colluvial wedges, no radiocarbon ages are available from this trench to constrain the ages of the paleoearthquakes.

3.3 Trench 3

Trench 3 (Photos 5, 6) crosses the antithetic fault on the northern edge of the graben toward the eastern tip of the 1981 rupture (Figs 12, 13, 18). At this location, the long-term net cumulative displacement of the fan surface measured by topographic profiling is ~2 m (in profile d, Fig. 14). The 1981 surface faulting produced ~0.6 m of max vertical displacement on the main scarp and an average 0.15 m free face on the antithetic, giving a net 1981 vertical displacement of ~0.45 m down to the north across the structure. Although located on a secondary structure (we may "miss" some events) this trench represents a highly favorable situation to recognize past surface faulting events because, on the one side, the antithetic scarp forms a morphologic trap to sediments washed on the fan surface and, on the other, the small throws on the antithetic fault may allow to "see" many events in a limited depth.

The trench exposes mainly alluvial sediments of variably sorted pebble conglomerates with a silty matrix, and scattered cobbles and boulders. These are topped by unit 5, which includes a reddish paleosol, and which marks the paleofan surface that is buried by colluvial and alluvial deposits on the downthrown block, and which is overlain by a thin, active, humic soil (unit 1) on the upthrown side of the antithetic fault (Fig. 18).



Fig. 18: Log of trench 3 from a 1:20 scale survey. .Colored lines highlight event horizons: I (green), J (blue), K (yellow), L(pink). Rectangles encloses cal ages of samples collected from both trench walls.



Photos 5 and 6: View of the east wall of trench 3 with the 1981 free-face in coincidence of the contact between.

The fault zone includes complex fault splays linked to the main antithetic fault, which steeply dips south (Photo 7). Some splays exhibit apparent reverse movement toward the free surface. Structural and stratigraphic relations were used to define four paleoearthquakes predating 1981 (Fig. 18).



Photo 7: View of the antithetic cumilative scarp near trench 3 and the 1981 free-face.

The **1981 event** ruptured the whole sequence in a deformation zone about 1 m wide including vertical offset and warping. A local maximum vertical offset of ~25 cm was measured. The penultimate **event I** is suggested by the occurrence of the scarp-derived colluvial wedge, unit 2, that on the east wall clearly overlies unit 3. **Event J** is defined by the significant faulting of unit 4 that remains after subtraction of the displacement produced by the younger events. The event horizon is well defined at the upward termination of the major apparently reverse fault splay south of the main fault (best seen in the east wall). Here, units 4 and older are faulted and downwarped against the fault. Once faulted, these units were "sealed" by the postevent unit 3, a yellowish to pale brown silty deposit of probable alluvial origin ponded against the scarp. **Event K** is responsible for the faulting of the complete alluvial fan package, up to the top of the red paleosol unit 5, and for the formation of the scarp-derived wedge of pebble conglomerates with a yellow to reddish silty matrix, unit 4. The contact between units 4 and 5 marks the event horizon. An older **event L** is hypothesized because of the additional displacement recorded by units 6 and 7 after removing the displacement produced by the younger events.

Four AMS ages are used to constrain the age of paleoearthquakes. **Event I** predates 1495-1680 A.D. (unit 2 - sample TR3-W06) and postdates 1265-1390 A.D. (unit 3 - sample TR3-E05), thus occurred between 1265 and 1680 A.D. **Events J and K** are both older than 1265-1390 A.D. (unit 3 - sample TR3-E05) and younger than 990-1165 A.D. (unit 5 - TR3-W04). Thus, these events occurred between 990 and 1390 A.D. and, because of their stratigraphic position, event J is expected to occur closer to the younger part of this range, whereas event K in the older part. **Event L** is younger than 670-875 A.D. (sample TR3-W02) and older than 990-1165 A.D. (unit 5 - TR3-W04). Is age lies between 670 and 1165 A.D., with the older part of this range preferred.

4. Results

Figure 19 presents a correlation of dated faulting events between the two trenches for which radiometric dates are available: trenches 1 and 3. The shaded areas encompass the possible ages (at the 2sigma interval) for each individual event, the darker zones highlights the preferred interval according to Pantosti et al., 1996 and Collier et al., 1998. Accepting that the error margins inherent to the dating technique do not allow proof of any correlation, the simplest interpretation of the data (requiring fewest paleoearthquakes) would correlate events 2 and 3 from trench 1 with events in trench 3. Event 2 in Trench 1 might correlate with event I or J in Trench 3. Assuming the first correlation to be correct, the penultimate event on the Skinos Fault is confined to the interval 1295-1680 A.D. It is possible that this event corresponds to the 1402 A.D. earthquake that produced important damage in the Corinth area accompanied by a tsunami (*Guidoboni and Comastri*, 2005). Radiometric dating allows event 3 of Trench 1 to be correlated with event J, K, or L in trench 3 (Fig. 19), with events K or L being the preferred options. If correct, this event is constrained to the time window 670-1015 A.D. Some events in Trench 3 are clearly not recognized in Trench 1.



Fig. 19: Correlation of Trench 1 and Trench 3 event ages. The age of the penultimate event and of three prior surface-faulting paleoearthquakes is constrained on the basis of dendrochronologically corrected ages of samples collected from Trenches 1 and 3. Black bars are the 2sigma age intervals for each sample (as labeled); arrows indicate whether the earthquake occurred after or prior to deposition of the sample. The shaded areas cover the possible age interval for each individual event, the darker zones locating the preferred part of the interval. Merging the results from both trenches, the simplest history constrains the age of the penultimate earthquake by correlation of event 2 in Trench 1 with event I in Trench 3. Event 3 in Trench 1 may match any of events J-L in Trench 3, but our preferred correlation is with either event K or L.

This implies either (1) that features describing discrete faulting events may have been lost by channel erosion or were too small to be recognized within the exposed Trench 1 section or (2) that one or more other en echelon segments, overlapping the segment revealed in Trench 1, may have taken up slip in intervening events. Either way, these results highlight the lateral variability of the fault structure and of the faulting record across the Bambakies Fan. We discard the hypothesis of a possible overinterpretation of trench 3 because, even though without any chronological constraint, also trench 2 support evidence for several surface faulting events since the redsoil on top of the fan surface started developing.

The data collected at this site allow also to estimate the following parameters describing the seismogenic activity of the Skinos fault:

- average vertical displacement in the range 0.4-1.2 m; these values are close to the 0.45-1.3 m 1981 displacements of the fan surface measured at the same localities;
- 2) average recurrence interval of 330 years derived from trench 3;
- 3) minimum average vertical slip rate of 1 mm/yr (from fan surface displacement), 0.7-2 mm/yr (derived from trench 1: 1-1.9 m of vertical displacement between 590-1015 A.D. and 1981) or 1.2-2.5 mm/yr (from offset of fan surface at trench 3 and dating from this trench); these compare with a long-term vertical slip rate of 1.2-2.3 mm/yr estimated on the basis of the present separation of the Megara pre-faulting surface located 8 km to the east.

Stop 3

Overview along the Skinos – Pisia road

Ioannis Papanikolaou and Gerald Roberts

The Schinos (or Skinos) segment continues offshore westwards from Alepochori village (Fig.20, Fig.21). Both Pisia and Skinos segments were activated in 1981 producing extensive surface ruptures and maximum displacements that reached of 150cm in Pisia and 100cm in Skinos (e.g. Jackson et al. 1982). There is some controversy as to whether the ruptures near Pisia and Skinos should be ascribed to the first or the second event of the sequence (Jackson et al. 1982; King et al. 1985; Taymaz et al. 1991; Abercrombie et al. 1995; Hubert et al. 1996), as both occurred at night and only a few hours apart. The maximum cumulative throw of both faults is observed between Skinos and Alepochori and estimated by several authors both onshore and offshore at about 2.5-3km (Myrianthis 1982, Perissoratis et al. 1986, Roberts 1996, Sakellariou et al. 2007) with approximately 1.5km on Skinos (Leeder et al. 2005). Roberts and Stewart (1994) have calculated also minimum 650m uplift on the Pisia footwall based on phreatic baroque dolomite cements.



Fig. 20: 3D view of the Bambakies fan and both Skinos and Pisia.



Fig. 21: 3D view of both Skinos and Pisia segments along the road from Skinos to Pisia. Both faults have a cumulative throw of 2.5-3km.

Stop 4

View point over marine terraces on the Perachora Peninsula.

Gerald Roberts and Ioannis Papanikolaou

A staircase of at least 9 marine terraces are developed on footwall of the Pisia-Skinos faults which are exposed on the south-facing shore of the Perachora Peninsula (Roberts et al. 2009; see also Turner et al. 2010 for an alternative view). Roberts et al. (2009) have U-Series coral dates from three of these which constrain the chronology of uplift. The coral dates are from specimens collected from shoreface sediments at the base of palaeo-sea-cliffs, just below notches carved into the limestones that are palaeoshorelines. Palaeoshorelines for glacio-eustatic sea-level highstands at 76 ka, 125 ka and 240 ka have been dated and mapped along strike. Modelling these with knowledge of sea-levels at other known highstands identifies the ages of other un-dated palaeoshorelines. Uplift rates decrease towards the western tip of the Pisia-Skinos Fault which lies a few kilometres west of Lake Vouliagmeni near the lighthouse at Heraion. These data have been interpreted to suggest that uplift rates increased by a factor of 3.2 ± 0.2 at 175 ± 75 ka, as explained below.

Holocene coastal notches and marine terraces.

We will walk along the beach from a car park behind a hotel to examine evidence of Holocene and Quaternary uplift rates (Figures 22 and 23).

First we will see Holocene coastal notches cut into a limestone rocky shoreline produced by physical erosion (wave action), chemical erosion (marine dissolution) and bio-erosion (boring/abrasion by sponges and bivalves that bore into limestone). Three notches, whose ages (< 6 ka, Pirazzoli et al. 1994) from ¹⁴C on attached organisms correlate with times of stable climate (Mayewski et al. 2004), are now uplifted. Although elsewhere it is clear that notches can form due to episodic uplift in earthquakes, the correlation with times of stable climate suggests that at this location individual notches formed when climatic conditions favoured intense erosion during progressive uplift of the shoreline (Cooper et al. 2007). The highest notch at ~3 m elevation implies an uplift rate of ~0.5 mm/yr.

Second we will see a location where corals associated with marine root-traces (very shallow marine) have been dated to 76 ka using U-Series determinations on corals. The corals and palaeoshoreline at ~12 m elevation imply ~42 m uplift since 76 ka (because sea-level was at -30 m at 76 ka; Imbrie et al. 1984; Siddall et al. 2003) implying ~0.5 mm/yr uplift rate.

Thus, uplift rates have been ~0.5 mm/yr throughout the Holocene and back to at least 76 ka at this position in the footwall of the Pisia-Skinos Fault. However, the sedimentary section seen at the next (nearby) location, with intercalated alluvial fans and marine sediments, imply that the uplift rate must have been lower in the early Quaternary, through comparison with the sea-level curve.

Alternative ideas about these notches and terraces are summarised in Turner et al. (2010).



Fig. 22: Interpetation of coastal notches adapted from Cooper et al. (2007)



Marine terraces and Holocene shorelines; uplift rate variation along the fault.

Fig. 23 : Overview of deformed palaeoshorelines on the Perachora peninsula (adapted from Roberts et al. 2009).

Outcrops of intercalated marine and alluvial fan sediments in an incised, but abandoned river gorge.

Exposures of intercalated lower and upper shoreface marine and alluvial fan sediments have been dated using U-Series on corals to reveal glacio-eustatic sea-level changes and fault slip-rates (Roberts et al. 2009; Fig. 24 and Fig. 25). Alluvial fan sediments, intercalated with marine sediments, are overlain by composite exposure/flooding surfaces that are encrusted with marine oysters associated with Lithophagid borings. Overlying marine deposits with coral colonies contain calcrete and hardground surfaces. Coral dates identify marine deposits from the 125 ka, 175 ka, 203 ka, 217 ka and 340 ka sea-level highstands. Intercalated calcretes and hardgrounds must therefore date from sea-level lowstands between these times. These periods of marine and continental sedimentation only correlate with the Quaternary sea-level curve if uplift increased by a factor of 3.2 ± 0.2 at 175 ± 75 ka. This is because the sub-aerial alluvial fan deposits near the base of the section could not have been formed if uplift rates had always been 0.5 mm/yr, because they would have been below sea-level before ~175 ka. Instead, uplift rates as low as 0.18 mm/yr prior to ~175 ka explain the episodes of sub-aerial exposure evidenced by the calcretes and alluvial fans.



Fig. 24 : Example of detailed mapping of palaeoshorelines (adapted from Roberts et al. 2009).







Fig.25 : Logs of intercalated sub-aerial and marine deposits (adapted from Roberts et al. 2009).
Stop 5

Wind gaps (palaeo-valleys) along the trace of the Pisia-Skinos Fault

Gerald Roberts and Ioannis Papanikolaou

The trace of the Pisia-Skinos Fault continues west of the town of Perachora. Surface ruptures to the 1981 earthquake passed through the town causing loss of life and severe damage (Jackson et al. 1982). We will examine the surface ruptures and also see a now-abandoned palaeo-valley (wind-gap) cut into the footwall of the fault (Morewood and Roberts 1999; Fig. 26, Fig. 27)). The Perachora drainage basin is strongly asymmetric. It is E-W elongated and controlled by the South Alkyonides fault system. Drainage formerly flowed down to the coast on the south of the Peninsula. However, lateral propagation of the surface trace of the Pisia-Skinos Fault has beheaded the drainage (Morewood and Roberts 1999). This caused abandonment of the palaeo-valley(s) which is now stranded and uplifted in the footwall, and divertion of the drainage to flow axially, parallel to the fault trace down to Lake Vouliagmeni. The lake (now connected to the sea through a human-cut canal) is itself ponded in by the footwall of the Pisia-Skinos Fault. The less pronounced footwall relief (a few tens of metres as opposed to the ~1000 metres seen at the Bambakies Fan) plus the relatively-small surface rupture (~15 cm coseismic throw see Fig 17 as opposed to ~1 m on the Bambakies Fan) mean that we are approaching the western tip of the Pisia-Skinos Fault. Work is currently underway to study the relationship between fault propagation, slip-rate changes and fault death in the Gulf of Corinth using dated marine terraces in the palaeo-valleys to date truncation of drainage (Fig. 18). Our hypothesis is that increase in the size of faults means they can accommodate more strain per unit time, so less active faults are needed through time if regional strain-rates remain constant - this drives displacement localisation (Cowie and Roberts 2001). Overall, it is suggested that when the surface expression of the fault achieved its present length, uplift rates increased by a factor of 3.2 ± 0.2 at 175 ± 75 ka, synchronous with the death of the across-strike Derveni fault at 382-112 ka (U-Series dates on carbonate cements coating the fault zone; Flotte et al. 2001). Before 175 ± 75 ka, both the Pisia-Skinos Fault and the Derveni Fault were active. After 175 ±75 ka, death of the Derveni fault and other faults may have transferred extra regional strain per unit time onto the Pisia-Skinos fault and perhaps other faults, forcing it to accelerate.





Fig. 26: View of the 1981 surface ruptures near the windgap locality. Broken roots and gaps are still visible 30 years after the events.



Fault evolution with varying sea-level

Fig. 27: Cartoon of the fault evolution on the Perachora Peninsula (adapted from Morewood and Roberts 1999).

Cowie and Roberts (2001) and Cowie et al. (2005), suggest that regional strain rates are maintained during the geometrical evolution of a fault system, so that the growth in size of some faults causes other faults in the system to cease activity (Fig.28). Specifically, following Cowie and Roberts (2001), the strain represented by a fault of length, L, and down-dip width, W, depends on the geometric moment, $M_a = dA$ where fault plane area A = LW, and W depends on the thickness of the faulted layer (Kostrov, 1974). According to the displacement-length scaling relationship $d_1 = \gamma L_1$, faults of length L_1 represent a strain M_{ρ} = yL_1^2W , whereas a larger structure that is developing through growth of the original smaller structures with a fully re-adjusted (self-similar) profile will eventually represent a strain $M_g = \gamma L_2^2 W$ (using $d_2 = \gamma L_2$). Note that $L_2 = NL_1$. It follows that the strain accommodated by the larger structure (with $d_2 = \gamma L_2$) is N times larger than that for the early faults, assuming W remains constant. In order for the regional strain rate to remain constant the spacing between those faults that remain active must change. In other words, some faults located across strike from faults that accelerate must cease or decrease in activity. We suggest that growth of faults in the Gulf of Corinth region may have produced an acceleration of the slip-rate on some faults in an attempt to achieve a future value of d_2 ; this redistributed regional strain rates causing some faults to cease or decrease their activity. In this scenario, early deformation may have occurred across the entire region. This was followed by growth and interaction between the faults, so that some faults, specifically those to the south of the Gulf of Corinth and those between the Gulf of Corinth and the Gulf of Evia, ceased or decreased in activity. If this is true, the velocity field measured with GPS for the Gulf of Corinth applies back to 175 ± 75 ka; the strain was more distributed prior to this time.



Fig. 28: Fault growth model. In order to maintain the d/L relationship as faults grow, fault plane area increases as earthquake occur causing lateral fault propagation. As fault plane are increases the fault accommodates more of the regional strain-rate, so through time, less active faults are needed. Some faults cease to be active driving displacement localisation. Faults that remain active shows increases in slip-rate through time. Adapted from Cowie and Roberts (2001).

Stop 6

View point from the escarpment along the south-dipping Loutraki Fault.

Gerald Roberts and Ioannis Papanikolaou

Looking south you can see the Corinth graben located between the south-dipping Loutraki Fault and the north dipping Kechriaie Fault. The Corinth canal is visible crossing the Isthmus of Corinth. The graben has developed since the early Pliocene (~5 Ma), but the exact age of initiation is uncertain. This extensional basin is located above the north-dipping Hellenic Subduction Zone and at the end of the North Anatolian Fault and is hence part of a back-arc extensional basin modified by strike-slip (Armijo et al. 1996). The Corinth Basin contains ~1 km of syn-rift sediment, but this increases to ~2 km plus ~1 km of bathymetry 30 kilometres to the west. Despite being in a graben, the Isthmus of Corinth is being uplifted in the footwall of the Pisia-Skinos Fault.



Fig. 29: a) View of the Loutraki postglacial scarp and of a probable relict windgap. b) View of the Loutraki fault plane and the deformed sediments.

The Gerania mountain (1351m) is a neotectonic horst that is constantly uplifted in Middle-Upper Pleistocene both from the South Alkyonides and the Loutraki faults located northwards and southwards respectively. A relict windgap can be probably traced towards the top of the mountain due to the ongoing uplift of both the south Alkyonides and the Loutraki faults (Fig.29). The Loutraki fault is the only fault in our fieldtrip that dips to the south. It has two parallel segments approximately 1.5km apart, one that bounds the Loutraki basin and it is also traced offshore (Lykousis et al.) and a second one at about 500m altitude that exhibits a postglacial scarp close to the Osios Potapios Monastery. The Loutraki Fault is active as it offsets a slope formed during periglacial activity in the last glacial maximum (~18 ka). It deforms sediments from the last glacial maximum exposed in a small quarry on the roadside. Slip in the Quaternary was ongoing, but intense erosion and sedimentation during glacial episodes would have outpaced scarp growth. Holocene slip and low sedimentation rates have produced a bedrock fault scarp that based on detailed scarp profiles is 8-9m high, implying it has a throw-rate of approximately 0.5mm/yr (Fig. 30 Papanikolaou and Roberts unpublished). Other well-known postglacial fault scarps in central and southern Greece are the Kaparelli and the Sparta faults both confirmed by cosmogenic isotope dating (Benedetti et al. 2002, 2003), the Delphi fault (e.g. Wiatr et al. 2011) and several scarps in Crete (Caputo et al. 1996).



Fig. 30: Distant and close up views and detailed topographic profiles across the post-glacial scarp of the Loutraki fault (Papanikolaou and Roberts unpublished data).

The Loutraki fault has no historical earthquakes associated with it, but Holocene slip implies a seismic hazard for the cities of Loutraki and Corinth that are built on unconsolidated coastal sediments. Despite the above the Loutraki fault has a significantly lower slip-rate (aprox. 1/4) compared to the Southern Alkyonides fault system and this can also partially explain why the area of Isthmus is uplifted.

Conclusions

The inventory of active faults around the Gulf of Corinth has changed through time with widespread extension across central Greece giving way to faulting that is mostly localised, at least in terms of strainrate, in the Gulf of Corinth and Gulf of Evia. This localisation appears to have occurred at 175 ± 75 ka for the location we have studied causing an increase in the fault slip-rate a factor of 3.2 ± 0.2 at this time. The key to recognition of this chronology is through the study of deformed palaeoshorelines and geomorphic features of known age. The ongoing challenge is to assess whether this chronology applied to the rest of the Gulf of Corinth.

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Field Trip Day 2

Ancient geographies and palaeotsunami hazard in the eastern Gulf of Corinth (Peloponnese, Greece)

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Fig. 1: Overview map (modified after Google Earth image, 2009)

Introduction

Tsunami hazard in the eastern Mediterranean and Greece

The eastern Mediterranean is characterized by the collision of the African and European lithospheric plates. Along the Hellenic Arc, the northward moving African Plate is being subducted under the Aegean microplate by rates up to 40 mm/a (Hollenstein et al., 2008). Seismic activity in the eastern Mediterranean thus belongs to the highest worldwide. Strong earthquakes occur with high frequency, often causing vertical displacements of the seafloor or submarine slides. These earthquakes are highly capable of triggering tsunamis (Papadopoulos et al., 2007; Papazachos & Dimitriu, 1991). Great water depths and additional short distances from shore to shore amplify the possibility of major tsunami events along Mediterranean coasts. Consequently, tsunami hazard in the eastern Mediterranean belongs to the highest in the world (Tselentis et al. 2010).



Fig. 2: Tsunamigenic zones in the eastern Mediterranean and epicentres of known tsunamigenic earthquakes (map modified after Soloviev, 1990; Papazachos & Dimitriu, 1991; Doutsos & Kokkalas, 2001). K₀ represents the (estimated) maximum tsunami intensity (Sieberg-Ambrasey scale), size of ovals reflects intensity estimates.

Many coastal areas in Greece are directly exposed to the Hellenic Arc, thus holding a particularly high risk of tsunami impact. Numerous historical accounts on tsunamis have been recorded all over Greece (e.g. Soloviev et al., 2000; Ambraseys & Synolakis, 2010; Fig. 2). Supplementary to archival studies, palaeotsunami research based on geo-scientific studies has been intensified during the past decades (Dominey-Howes et al., 2006; Dawson & Stewart, 2007; Mastronuzzi et al. 2010, Vött et al., 2010a, 2010b) to evaluate and better understand the influence of tsunami impact on coastal geomorphology and coastal evolution in Greece.

Earthquakes and tsunamis in the Gulf of Corinth

Throughout the Mediterranean, the Gulf of Corinth belongs to one of the seismically most active regions. On-going continental rifting with extension rates of 5 to 15 mm/yr has created a half-graben structure that is surrounded by numerous active on-shore and submarine faults (Papazachos & Dimitriu, 1991; Sachpazi et al., 2003; Avallone et al., 2004). Strong earthquakes up to M = 7 were observed in the gulf. For the past six centuries, 30 earthquakes with magnitudes $M \ge 6$ were recorded (Fig. 3). The average return rate for earthquakes of M = 6 is estimated to be 20 years (Papadopoulos, 2003; Stefatos et al.; 2006). Although it represents a rather small and enclosed basin, the Gulf of Corinth thus holds a comparably high potential for tsunami events. Numerous tsunami catalogues report on historical tsunamis that occurred within the Gulf of Corinth (Soloviev, et al. 2000; Papadopoulos, 2003; Ambraseys & Synolakis, 2010). Six events were recorded solely for the 20th century (Tab. 1); the most destructive occurred in 1963 near Aeghio (Papadopoulos, 2003).



Fig. 3: Major onshore faults, epicentres of earthquakes M > 6 and recorded tsunami events for the last 600 years (map modified after Sachpazi et al., 2003, compilation of epicentres after Stefatos et al., 2006 and tsunami events after Papadopoulos, 2003).

ID	Year	Region	Description	ті	TR
1	- 373	WCG	Destructive sea inundation	5	4
2	1402	CCG	Destructive inundation	4	3
3	1742	WCG	Flooding in Vostiva	3	3
4	1748	WCG	Destructive waves in Aeghio	4	4
5	1769	ECG	Sea flowed back in Desfina	2	3
6	1794	WCG	Strong inundation in Galaxidi	3	3
7	1817	WCG	Destructive waves in Aeghio	4	4
8	1861	WCG	Damaging waves in Aeghio	3+	4
9	1887	WCG	Sea-wave in Galaxidi	2+	4
10	1888	WCG	Sea-wave in Galaxidi	2	4
11	1898	CG	Sea-waves	2+	3
12	1963	WCG	Destructive waves in Aeghio	4	4
13	1965	WCG	Strong wave in Eratini	3	4
14	1981	ECG	Weak wave	2	4
15	1984	WCG	Strong wave in Sergoula	3	4
16	1995	WCG	Sea-wave in Eratini	2+	4
17	1996	WCG	Sea-wave in Aeghio	3+	4

Tab. 1: Tsunami catalogue for the Gulf of Corinth (modified after Papadopoulos, 2003). ID – identification number, WCG – West Corinth Gulf, ECG – East Corinth Gulf, CCG – Central Corinth Gulf, CG – Corinth Gulf, TI – tsunami intensity in Ambrasey (1962) scale, 1: very light - 6: disastrous, TR – reliability of tsunami event, 1: improbable tsunami - 4: definite tsunami.

Tsunami hazard in the Gulf of Corinth is controlled by three major factors. Along the faults bordering the gulf shallow earthquakes are generated, often accompanied by co-seismic displacements of the seafloor (Tinti et al., 2007). Based on the analysis of historical data, earthquakes with magnitudes M \geq 5.5 occurring near-shore or offshore seem to be most tsunamigenic (Papadopoulos, 2003). A prevailing fault orientation parallel to the coastline further allows easy wave propagation in northern or southern direction. With water depths of maximum 900 m, steep submarine slopes and a narrow shelf, the bathymetry of the gulf additionally enhances the development of strong tsunami waves (Papadopoulos, 2003; Stefatos et al., 2006;). Due to high sedimentation rates and steep slopes, coastal landslides or submarine slides are often triggered by a seismic event and must thus be considered as another significant origin for tsunamis. Regarding the spatial distribution of tsunami events throughout the Gulf of Corinth, an eastward decrease in tsunami frequency as well as intensity may be derived from the recorded tsunamis (Hasiotis et al., 2002; Papadopoulos, 2003).

Stop 1

Acrocorinth Coordinates: N 37°52'36", E 022°52'27"

Topographical and geological overview

Modern Corinth as well as the archaeological site of ancient Corinth are situated at the south-eastern margin of the Gulf of Corinth, the so called Lechaion Gulf. The modern city is located at the Isthmus of Corinth connecting the Peloponnese with mainland Greece. The present day topography of the wider Corinth area, the Corinthia, is largely the product of an on-going regional uplift affecting the northern Peloponnese and considerable eustatic sea level changes throughout the last 0.5 Ma (Hayward, 2003).

The north-eastern Corinthia is dominated by a coastal plain which extends up to 4 km inland and consists of unconsolidated Quaternary deposits (IGME, 1972). Along the shore, recent beach deposits mainly consist of sand and gravel. At various places, consolidated sediments form beachrock complexes. The coastal plain is bordered by a sequence of marine terraces forming a south-westward rising staircase with scarps running predominantly parallel to the coastline of the Lechaion Gulf (Fig. 4). With heights up to 820 m (south-east of Xylokastro) the terraces dominate the topography of the south-western Corinthia and belong to the highest uplifted marine terraces worldwide. They consist of yellow to white marls of Pliocene to Pleistocene age formed during sea level high stands and subject to subsequent tectonic uplift. The marls are unconformably overlain by marine and near-shore conglomerates and sands, again covered by alluvial and colluvial deposits of Holocene age. Macrofaunal remains including *Strombus bubonius* classify the conglomerates as Pleistocene (IGME, 1972; Keraudren & Sorel, 1987). The terraces are dissected by gullies that are orientated towards the coastline. In wide areas, the porous Pleistocene deposits serve as a protective cap against further erosion of the underlying marls.



Fig. 4: Simplified geological map of the Corinthia (modified after IGME 1972, sheet Korinthos).

Due to the fact that the western terraces have been uplifted to higher elevations, increased erosion has led to an on-going removal of the protective cap and exposure of the Pliocene marls. In those areas, the development of badlands can be observed (Hayward, 2003).

Another prominent feature in the topography of the Corinthia is the hill of Acrocorinth with its summit at 574 m above sea level. It mainly consists of limestone from the Middle Jurassic and sporadic layers of shale (IGME, 1972; Higgins & Higgins, 1996). At the entrance to the archaeological site a thrust fault can be observed, where the limestone overlies thick shale units that build up the western slopes of Acrocorinth (IGME, 1972; Fig. 5). The western flank of Acrocorinth is connected to the surrounding hills by the low shale ridge whereas the northern and eastern flanks are characterized by steep cliff-like slopes. The hilltop of Acrocorinth comprises an eastern and a western summit with the maximum elevation occurring in the eastern part and a central saddle (Carpenter & Bon, 1936).



Fig. 5: (a) Gullies intersecting the terraces of Corinth. View from Acrocorinth towards the west. (b) Thrust fault separating Jurrasic limestone from shale visible at the northwestern entrance to the fortification complex at Acrocorinth. (c) Detailed view of shale.

The fortifications of Acrocorinth

In antiquity, Acrocorinth served as a refuge for the population of ancient Corinth. Overlooking the Isthmus of Corinth and controlling the passage between the Peloponnese and mainland Greece, it was of great strategical importance. Due to its geomorphology, the fortress was quite easy to defend. A further fortification of the site by walls as well as a continuous water supply by the Upper Peirene fountain made Acrocorinth a secure stronghold.

Today, various fortifications are still preserved, as best visible from satellite images (Fig. 6). A closed fortification wall surrounds the heights and the saddle. In antiquity a gate and adjacent towers were built to protect the northwestern entrance at the lower level of the saddle. The fortification wall visible today is predominantly of medieval age, but follows the circuit of the ancient wall. A further fortification of the entrance was conducted during medieval times comprising a triple line of walls as well as an outer and inner gate (Carpenter & Bon, 1936; Scranton, 1960). From the 11th century onwards, the fortress at Acrocorinth was under Frankish, Byzantine, Venetian and Turkish control, as still visible from numerous building remains. It remained occupied by the Turks until the Greek War of Independence in the early 19th century (Scranton, 1960). The excavation and reconstruction of the site began in 1929, conducted by the American School of Classical Studies at Athens.



Fig. 6: (a) Fortifications of Acrocorinth as seen from the air (modified after Google Earth image, 2009). (b) Acrocorinth seen from the coastal plain looking south. (c) Fortified entrance with triple walls and towers.

Stop 2

The archaeological site of Archaia Korinthos

Coordinates: N 37°54'20", E 022°52'47"

A short history of ancient Corinth

The site of ancient Corinth – located on top of an uplifted marine terrace some 3 km distant from the Gulf of Lechaion at 60 m above present sea level – was more or less continuously settled since 3000 BC as revealed by archaeological evidence (Keraudren & Sorel, 1987). Around 900 BC, the city came under Dorian domination and continuously expanded. At that time, Corinth already held a significant commercial as well as military power. This was mostly based on its naval strength which is documented by the foundation of the Corinthian colony Corcyra (Corfu) during the 8th century BC. Also, Corinth was involved in the earliest recorded naval battle among Greek opponents carried out between the Corinthians and Corcyraeans in the early 6th century BC. The famous temple of Apollo is associated to this very era. From the late 6th century BC, especially under the tyrant Periander, the commercial and military expansion of the city was further accelerated. During the 5th and 4th century BC, ancient Corinth represented one of the wealthiest cities in the Greek world and mother city of numerous coastal colonies founded in the eastern Mediterranean (Graham, 1964). However, it was more and more affected by the constantly growing competitive Athenian sea power. Within the framework of repeatedly shifting alliances with Sparta and Athens, the Corinthians participated in the Peloponnesian as well as the Corinthian wars. During the late 4th and early 3rd centuries BC the city was held by the Macedonians (Polyb. Hist. 5.2.4-11 after Paton, 1923). Due to increasing conflicts with Rome, Corinth was destroyed by the Roman consul Mummius in 146 BC and re-founded by Julius Caesar some 100 years later at 44 BC (Paus. 2.1.1 after Jones, 1918; Fowler & Stillwell, 1932). For the period of four subsequent centuries the city again expanded and regained its former wealth as documented by numerous representative buildings from the 2nd century AD. In 375 AD, Corinth was affected by a series of strong earthquakes, in 395 AD, the city was invaded by the Goths, both events causing extensive damage. Throughout the 6th century, a series of strong earthquakes affected Corinth and led to its final destruction in 521 or 551 AD (Scranton, 1960; Pallas, 1990).

In medieval times Corinth was only of little significance. From the 11th century AD onward, the city was under the rule of Frankish, Byzantine, Ottoman or Venetian powers. The modern city of Corinth, located in the vicinity of its ancient predecessor, developed from a village founded under Turkish control. It was destroyed by an earthquake in the 19th century AD (Scranton, 1960). Later, it was refounded at its present day location close to the Corinth Canal. At the end of the 19th century, the site of the ancient city was indicated by the ruins of the Apollo temple (Fowler & Stillwell, 1932).

The archaeological site

Excavations at ancient Corinth started more than a century ago, in 1896, conducted by the American School of Classical Studies at Athens (ASCSA). Firstly focused to the area around Agora, on-going excavations revealed the complex structure of ancient Corinth and its surrounding area. The remains, visible in the archaeological site, are mostly of Roman age (Fig. 7), but evidence of the Greek city is still preserved underneath (Scranton, 1960).

The Lechaion Road

East of the Apollo Temple, a natural depression trends in N-S direction. During Greek and Roman times, it was artificially widened and gave access to the Agora. Through the depression, a wide road heading northward connected the ancient city with its harbour site Lechaion at the shore of the Gulf of Corinth, situated about 3 km away from the city. As Pausanias notes, *"on leaving the market place along the road to Lechaeum you come to a gateway"* (Paus. 2.3.2 after Jones, 1918). This Roman gateway, the so-called Propylea, directly approached the north-eastern corner of the Agora. According to Pausanias, the Lechaion Road was the main route to the harbour *("Proceeding on the direct road to Lechaeum [...]"*, Paus. 2.3.4 after Jones, 1918).

Excavated only in the area close to the Agora (Fig. 8), the course of the road could be followed by random pits up to a few hundred meters towards the north. It is paved with large limestone slabs retrieved from local quarries at Acrocorinth. Either side of the road was boarded by a sidewalk. All



along its way to Lechaion, the entire road was protected by accompanying fortification walls connecting the important harbour site to the ancient city itself.

Fig. 7: Excavated remains of ancient Corinth as visible today (modified after Scranton 1960).

There is evidence that the road had steps at certain intervals to overcome passages with steep slopes. Therefore it could not be used by vehicles (Fowler & Stillwell, 1932; Scranton, 1960).



Fig. 8: View of the Lechaion Road towards the north.

Stop 3

Lechaion, the ancient harbour of Corinth

Coordinates: N 37°55'57'' E 022°53'11''

Lechaion - a brief history of the harbour

The archaeological site of Lechaion, located at the southeastern shore of the Gulf of Corinth, represents the western harbour of ancient Corinth. The harbour foundation was most probably associated to the expansion of Corinthian military and trading activities in the late 7th century BC and is ascribed to Periander, second tyrant of Corinth. Due to its favourable location at the Isthmus of Corinth, the harbour was probably in use for almost a millennium (Sanders & Whitebread, 1990; Rothaus, 1995; Stiros et al., 1996). Ancient Corinth also possessed an eastern harbour at the shore of the Saronic Gulf, called Kenchreai, situated some 10 km to the east of the ancient city.

During the Corinthian War in the early 4th century BC, Lechaion served as a naval base for the Corinthian or Spartan fleets. According to Xenophon, the harbour held a complex infrastructure including dock houses, ramparts and fortification walls. These walls, also connected the harbour to the city of Corinth. Lechaion thus belonged to the *epineon*-type of harbour (Xen. Hell. 4.4.6-8 after Brownson, 1918; Lehmann-Hartleben 1923). Around 200 BC, the harbour became the main base for the Macedonian fleet under Philipp V. As Polybius notes, six thousand Macedonian and Achaean soldiers and twelve hundred mercenaries set out from Lechaion, which allows the estimation of at least 40 triremes assembled at the harbour (Polyb. Hist. 5.2.11 after Paton, 1923; Dahlheim 1997). With the Roman devastation of Corinth in 146 BC, Lechaion was most probably not destroyed but abandoned for a century. It regained its former importance after the re-colonisation of Corinth under Julius Caesar in 44 BC.

In Roman times, Lechaion underwent at least two phases of reconstruction throughout the 1st and 4th centuries AD. The harbour most likely served for economic rather than military purposes (Strab. Geogr. 8.6.23 after Hamilton and Falconer, 1903; Kent, 1966). The final abandonment of Lechaion is associated with the destruction of Corinth by a series of strong earthquakes in 521 or 551 AD. Though occasionally re-used in medieval times, the harbour never regained its former importance (Rothaus, 1995). Historic accounts from travellers in the 19th century AD still refer to Lechaion as a "*cove*" or "*lagoon*" close to modern Corinth but describe the site as "*desolated*" and "*deserted*" (Blaquiere, 1825; Leake, 1830).

Visible remains of harbour works

Until today, the archaeological site of Lechaion has remained mostly unexcavated. Nevertheless, certain harbour installations are still visible and allow a rough estimation of the former harbour structure. According to Paris (1915), Lechaion was separated into an outer and inner harbour basin (Fig. 9), of which the latter seems to be quite well preserved. An outer basin is indicated by two large moles running perpendicular to the present coastline (Fig. 10a). Further east, remains of moles define the entrance to the inner harbour, a channel about 150 m long and 14 m wide (Figs. 10b,c).



A₁₋₂ - outer harbour basins, M₁₋₂ - remains of moles, J₁ - entrance moles to inner harbour, P₁₋₂ - entrance channels to inner harbour, D₁₋₄ - inner harbour basins, E₁₋₂ - basins for navigation, N₁ - narrow passage between inner basins, Q₁₋₄ - quay remains, F₁ - foundation

Fig. 9: Overview plan of the present day Lechaion harbour basin and visible remains of harbour installations (modified after Paris 1915).

The elongated inner harbour basin visible today can be subdivided into two main basins probably used for navigation and four adjacent basins interpreted as sites for anchorage (Paris, 1915; Lehmann-Hartleben, 1923). In some parts, ancient quay walls are well preserved (Fig. 10d). Additionally, remains of a harbour monument or small building most probably of Roman age are still present in the central harbour basin (Shaw, 1969; Rothaus, 1995).

The present day topography of the harbour site is dominated by large mounds of sediment, generally associated to dredging activity or referred to as natural dunes (Frazer, 1965; Rothaus, 1995; Stiros et al., 1996). Up to 16 m high, these mounds are located in the eastern harbour area, situated to both sides of the entrance channel.

The Lechaion basilica

In the western area of the Lechaion harbour site, remains of an ancient church constitute a remarkable archaeological find (Fig. 11). Excavations accomplished by D. Pallas between 1955 and 965 revealed the well preserved foundation of an Early Christian basilica (Pallas, 1960). With a total length of 186 m this basilica represents the largest known sacred building from its period and is comparable to St. Peter in Rome. The basilica is situated between the inner harbour basin and the



Fig. 10: Visible remains of harbour installations. (a) Outer moles M_1 (background) and M_2 (front), view to the NW. (b) Mole J_1 at the entrance to the inner harbour basin, view to the NNE. (c) Entrance channel P_1 to the inner harbour basin for navigation E_1 and adjacent sediment mounds, view to the S. (d) Remains of ancient quay Q_2 , view to the S. (e) Foundation F_1 at inner harbour basin D_3 , view to the N. For details and indices see

present beach some 100 m distant from the present shore and was thus clearly visible from both land and sea (Krautheimer, 1989; Rothaus, 1995).

The obviously exceptional location and dimensions of the building must be considered against the historical background of the development of Christianity in Greece. From the 1st century AD,



Fig. 11: Plan of the Christian basilica at Lechaion (after Pallas 1960).

Christian congregations first emerged in the Greek society, but it is only from the 5th century AD onwards that Greece experienced an increasing popularity of Christianity (Krautheimer, 1989). Christian influence in Greek cities especially developed in coastal areas, for seafarer came in close contact with the capital of Constantinople. Large churches were therefore often built in coastal cities such as Corinth. Here, at least five basilicas were constructed in or close to the ancient city, including the Lechaion basilica (Krautheimer, 1989; Scranton, 1960).

Columns encountered during excavation works indicate a 5th century age for the Lechaion basilica, as must be assumed from finds of coins associated to the construction of the foundation and inner pavement. The construction is thus assumed to have started in the middle of the 5th century AD and most probably continued until the early 6th century AD (Pallas, 1960; Krautheimer, 1989; Rothaus, 1995).

Tectonic uplift in the Lechaion area

The tectonic evolution of the Gulf of Corinth is characterized by horizontal as well as enormous vertical movements. While the northern coast is dominated by subsidence, the southern coast of the gulf has been subject to ongoing major tectonic uplift since mid-Pleistocene times. However, the uplift shows a certain regional variability. Maximum rates occur in the central part of the gulf and decrease in eastern and western direction. Thus, the amplitude of the Quaternary uplift lies in between 900 m and 1600 m along the central part of the southern coast. Holocene shorelines, well preserved above the present day sea level, allow the estimation of an uplift rate of approximately 3 mm/year. In the Corinth area, an amplitude of maximum 880 m indicates lower uplift rates (Pirazzoli et al., 2004).

The increased uplift in western direction is also reflected in the terraces of Corinth. Accordingly, the New Corinth terrace lies at 30 m a. s. l. close to modern Corinth, but increases to 75 m a.s.l. near Vocha and reaches 160 m a.s.l. near Xylokastro. The older Ancient Corinth level reaches from 60 m a.s.l. near Ancient Corinth to 145 m a.s.l. near Vocha to 360 m a.s.l. near Xylokastro (see Fig. 4). The average uplift is estimated at 1.5 mm/year. Higher average rates are to be expected for the Xylokastro area, while lower average rates occur around Corinth (Keraudren & Sorel, 1987).

At Lechaion, evidence of tectonic uplift can be observed along different harbour installations. At the entrance to the inner harbour, for example, borings of marine molluscs are exposed at > 1 m above present sea level. In some holes, articulated bivalves of *Lithophaga lithophaga* are still preserved. Thus, a rapid uplift of a certain rate must be assumed (Pirazzoli et al., 1996). Otherwise, the shells would have been destroyed by erosion. Age determinations of single specimens by radiocarbon dating range from 600 cal BC to 50 cal BC with a clustering around 340 cal BC +/- 100-150 indicating a single uplift event between the 5th and 3rd centuries BC (Stiros et al., 1996). Pirazzoli et al. (1996) specify at least 1.2 m a.s.l. of uplift, also indicated by marine organisms. However, radiocarbon dates suggest not one, but two uplifting events, one at 340-130 cal BC [?] and another at 442-657 cal AD. Tectonic uplift is commonly referred to as an explanation for the abandonment of the harbour (Stiros et al., 1996).

Tsunami impact on the Lechaion harbour site

As previous studies have shown, tsunami impact may leave different sedimentary signatures comprising (i) dislocated blocks and boulders, (ii) layers of allochthonous marine sediments intersecting quiescent near-shore environments such as lagoons, swamps or freshwater lakes, (iii) geo-archaeological destruction layers consisting of allochthonous marine sediments intermingled with terrigenous material and cultural debris, and (iv) beachrock-type calcarenitic tsunamites (Vött & May, 2009, Vött et al. 2010a). For the preservation of palaeotsunami signatures, low-energy environments such as ancient harbour basins represent excellent geological archives. Thus, geomorphological, sedimentological, geochemical and geophysical studies were carried out in the inner harbour basin of Lechaion.

Vibracorings drilled in the Lechaion harbour have revealed shallow marine (pre-harbour) deposits overlain by mud accumulated in the quiescent lagoonal environment. Early harbour deposits are covered by limnic to terrestrial sediments that were deposited when the harbour was already out of use. The Lechaion harbour stratigraphical record is repeatedly interrupted by layers of allochthonous coarse-grained sediment and shell debris, partly characterized by fining upward sequences. These ex situ-layers were found up to 450 m inland from the present shoreline. Erosional unconformities at their base as well as immediately re-established quiescent conditions on top, reflected by autochthonous mud, document short term high-energy interference of the Lechaion harbour site. With regard to the overall dimensions of the event layers and the geographical position of the study area, storms can be excluded as triggering factors. Both geomorphological and sedimentary fingerprints of the high-energy event layers rather indicate a tsunamigenic origin. In the Lechaion harbour area, we found three distinct tsunami layers. The geochronological framework based on radiocarbon dating and age determination of diagnostic ceramic fragments suggests tsunami impacts around 760 cal BC, 50 cal AD and during the 6th century AD (Hadler et al., 2011).

Large sediment mounds adjacent to the inner harbour basin dominate the present day topography of the Lechaion harbour site the origin of which is explained by natural dune formation or sediment obtained from dredging activities (see above). Mound sediments mainly consist of sand and gravel intermingled with abundant ceramic fragments and marine macrofossils, a fact that principally excludes natural dune formation. However, dredging usually aimed to remove predominantly finegrained homogeneous harbour sediments. Coarse-grained sediments of the type encountered in the quiescent Lechaion harbour basin thus do also not correspond to typical dredge material. As the coarse-grained deposits are identical to the youngest tsunamite encountered in adjacent trenches, we assume that the mounds are built up of tsunami deposits that were removed from parts of the pre-event harbour basin soon after a strong tsunami event.

High-energy sediment input to the harbour site must also be assumed with regard to the buried ruins of the 5th century AD Christian basilica. The basilica's entire foundation as well as the adjacent harbour area are completely covered by a sediment layer up to 2 m thick (Fig. 12).



Fig. 12: Christian basilica at Lechaion. (a) Overview showing foundation of basilica as seen from its eastern end, close to the inner harbour basin. (b) Height difference between the present day ground surface (foreground) and the pavement of the baptistery. (c) Typical column dating to the middle of the 5th century AD. (d) Wall fragments most probably displaced due to collapse of the building during the 551 AD earthquake. (e/f) Embedding of the basilicas foundation. Note the considerable height difference (Δh) between the foundation level and the present day ground surface. (g) Accumulated sediments consist of sandy/silty matrix including with gravel and ceramic fragments.

Like the dredge mounds, the sediment shows a sandy matrix including gravel, ceramic fragments and marine macrofossils. A former door lintel later used as a threshold indicates that the walking level considerably changed by massive sediment accumulation during which large parts of the church were destroyed. The results of our study suggest a tsunamigenic cause for both sediment accumulation and destruction. The topography of the coastal plain as well as the composition of the high-energy sediments also exclude a colluvial, fluvial, mass denudative or man-made burial of the site.

Earth resistivity transects carried out across the ancient harbour basin revealed a sharp boundary between the allochthonous coarse-grained high-energy deposits and the underlying autochthonous fine-grained harbour sediments (Fig. 13). The event layer can be traced up to 450 m inland showing a clear thinning landward structure. Ground penetrating radar profiles revealed channel-like structures at the base of the uppermost tsunami layer orientated in a right angle towards the present coastline and documenting strong linear erosion. Similar structures were recently produced by the February 2010 Chile tsunami by strong backflow (Bahlburg & Spiske, 2010).



Fig. 13: Earth resistivity transect LEC ERT 3 measured in the inner harbour area. The contrast between autochthonous harbour sediments (green to blue) and allochthonous tsunami deposits (red to purple) is clearly visible.

Our results revealed repeated strong tsunami impact for the Lechaion harbour site and adjacent coastal areas. The present day topography of the harbour and the geomorphology of the inner harbour basin are most probably due to a partial re-excavation of the harbour basin during medieval or pre-modern times.

A timeframe for the final burial of the Lechaion harbour site is given by archaeological evidence. According to Rothaus (1995), only few ceramic fragments younger than late Roman to early Byzantine times were found in the vicinity of the harbour basin. Thus, a sudden abandonment of the site must be assumed for the 6th century AD. The construction period of the basilica itself, embedded in tsunamigenic sediments, also gives the 6th century AD as a *terminus post quem* for the event. Strong tsunami landfall in the 6th century AD must thus be regarded as the most probable cause for the final destruction and abandonment of the Lechaion harbour complex. The event seems to be related to the 521 or 551 AD earthquakes which are reported by historic accounts to have destroyed ancient Corinth.

Stop 4

Beachrock-type tsunami deposits at the Corinth Canal

Coordinates: N 37°57'05'' E 022°57'38''

The "diolkos" – an ancient slipway for ships across the Corinth Isthmus

In antiquity, ships travelling between the Corinthian and the Saronic Gulfs had to circumnavigate the Peloponnese on a long and difficult journey around the dangerous Cape Malea. Therefore, various attempts were made throughout history to cut a channel through the Isthmus of Corinth. But it was not until 1893 that the project was finally realized by a joint venture of French and Hungarian enterprises. The Corinth Canal allowed a short and time-saving passage for ships from the Gulf of Corinth to the Saronic Gulf (Lewis, 2001). Today, the canal is too small for larger cargo ships and is primarily of touristic interest.

In antiquity a slipway, the so called " $\delta i o \lambda \kappa o \varsigma$ ", already led across the narrowest part of the isthmus connecting the Lechaion Gulf with the Saronic Gulf. Based on archaeological evidence, the diolkos most probably dates to the late 7th or 6th century BC possibly built under the rule of the tyrant Periander (Lewis, 2001). At several occasions throughout history ships were towed across the diolkos (MacDonald, 1986; Lewis, 2001). During the Peloponnesian War in 428 BC, the Spartans had already prepared to haul their ships across the isthmus but then had to draw back against the Athenians (Thuk. 3.15 after Landmann, 2010). It must be assumed, though, that the transport of ships was generally limited to warships, for they were usually of less weight and narrower shape than heavy and broad merchant ships.

Based on literary evidence, the diolkos was used at least between the 5th century BC and the mid-1st century AD. There is no historical record of the diolkos after 67 AD, when it was most probably truncated by Nero's attempt to cut a channel across the isthmus (Lewis, 2001). Today, remains of the ancient slipway are still visible at the western entrance to the Corinth Canal. Here, a pavement of large, squared blocks up to 1.6 m length and 1.1 m width is preserved, partly lying below sea level. This pavement is covered by a calcified layer of sand and gravel, postdating the ancient diolkos (Fowler & Stillwell, 1932).

Beachrock-type tsunamigenic deposits

Along the Corinthian coastline, extensive beachrock formations dominate the present coastal geomorphology from Lechaion to Loutraki (Fig. 14). As recently described by Vött et al. (2010a), beachrock deposits may also be interpreted as sediments of tsunamigenic origin that were subject to post-depositional calcification and are thus preserved along present day coastlines.

At the Corinth Canal, some 7 km distant from Lechaion, the beachrock complex extends up to 300 m inland, partly covering the ancient diolkos. The beach along the present shore between Lechaion and Loutraki, however, only reaches an average width of 15 - 20 m and does not exceed a maximum width of 30 m.



Fig. 14: Beachrock along the Corinthian coast. (a) Extensive beachrock at the Corinth Canal. (b) Ancient diolkos covered by beachrock. (c) Internal structure of beachrock showing lamination and multiple fining upward sequences. (d) Beachrock outcrops along the Loutraki beach.

Thus, even in low lying coastal stretches the prevailing coastal dynamics are obviously not capable of transporting sediments as far inland as indicated by the beachrock. The encountered beachrock consists of clearly laminated sediments showing multiple fining upward sequences from gravel to fine sand. A landward orientated imbrication of the gravel components can be observed, thus documenting landward flow dynamics. Both sedimentary characteristics are not typical of a littoral environment. The beachrock therefore must not be regarded as lithified beach but rather represents beach sediments accumulated far inland during strong tsunamigenic inundation of the Corinth coastal area.

Stop 5 Submerged harbor and temple at Kechriaie

Ioannis Papanikolaou and Gerald Roberts

Ancient Corinth was a major naval force, a major trade centre and exporter of black-figure pottery to city-states from 600 BC. It rivaled Athens in wealth, based on the Isthmian traffic and trade. This trade was facilitated by the two main ports of the town; the Kechriaie port located in the Saronikos Gulf and the Lechaion port located in the Corinth Gulf. Ships were transported overland between both harbors through Diolkos a limestone paved trackway approximately 8km long. Kechriaie (or Kechries or Kechriai or Kenchriaie in modern Greek : Κεχριές) is also famous because Apostle Paul visited the harbor at least three times having a prolonged stay during his third visit and mentioned it twice to his epistles to the Corinthians.



Fig.1: View of the submerged Kechriaie harbor.

The submerged harbor and the temple of Isidas that are now up to 2 m below sea level indicate that seismic activity, has led to the subsidence of the coastline since ancient times (Fig.1, Noller et al. 1997). Subsidence should occur no earlier than 80 A.D. so Kechriae is considered one of the clearest archaeological indicators of earthquake destruction and co-seismic subsidence and illustrates the gap between archaeological science and paleoseismic data (Rothaus et al. 2008). Herein it has to be mentioned that the nearby area of Isthmus is constantly uplifted approximately 0.3mm/yr over the last 200kyrs (Collier et al. 1992), therefore no regional subsidence is justified, other than seismic events. The Kechriaie harbor is surrounded by three major seismic sources (the Kechriai fault, The Loutraki fault and the Agios Vassileios) that can all produce subsidence to the temple since it is located on their hangingwall (Fig.2). Two of these seismic sources are distant (the Loutraki fault at 11 km and the Agios Vassileios at 8 km) and one is proximal (the Kecrhiaie), but they all have low sliprates, not exceeding 0.5mm/yr (Collier and Thompson 1991, Roberts and Jackson 1991, Papanastassiou and Gaki-Papanastassiou 1994, Papanikolaou et al. 1989, Papanikolaou et al. 1996, Goldsworthy and Jackson 2001, Rondoyanni et al. 2008, Zygouri et al. 2008). A smaller active fault towards the eastern tip of the Corinth Canal near Kalamaki has a very low slip-rate and is too short to significantly impact on the Kechriai harbor (Papanikolaou et al. 2011).



Fig.2: Simplified map of the major seismic sources surrounding the Kechriaie submerged harbor.

The harbor lies in the immediate hangingwall of the Kecrhiaie fault, implying that a significant subsidence is expected to occur during rupturing of this fault. Whether submergence was instantaneous or gradual remains an open question. However, the almost 2 meters of subsidence most probably could not have not occurred by a single event, but resulted by the cumulative displacement of multiple events. Moreover, mosaics and paved floors were found level and intact, so no evidence for liquefaction phenomena were traced, implying that subsidence is totally fault-controlled (Rothaus et al. 2008).



Fig.3: 3D view of the submerged harbor in Kechriaie, located in the immediate hangingwall of the Kechriaie fault.

Kechriai is an E-W trending normal fault that downthrows towards the north (Fig.3). It propagates also offshore for at least 10 km in the Saronikos Gulf where lithoseismic profiles show offset of the sediments (Papanikolaou et al. 1988, 1989). It is segmented and exerts an impact in the topography, but no postglacial scarp is revealed like the one in the Loutraki fault (see Fieldtrip 1 Stop 6), implying that it has a low slip-rate (<0.3mm/yr). In 1858, a M=6.5 earthquake destroyed Corinth (X intensity).

The area of damages followed an E-W direction, which is parallel to the fault strike values of the area, with its macroseismic epicenter recorded in Acrocorinth (Papazachos and Papazachou 1997). In 1928, another earthquake of M=6.3 damaged Corinth (IX intensity), during which 20 people died and 3.000 houses were destroyed in Corinth and Loutraki (Papazachos and Papazachou 1997). It produced several secondary surface ruptures over the entire area and Kalamaki village that lies on the immediate hangingwall of the Kechriai fault was entirely destroyed. Therefore, the Kechriai fault may have hosted this earthquake. In addition, southwards from the Kechriaie fault the Agios Vassilios (or Athikia) is major E-W trending normal fault zone approximately 38km long, that downthrows towards the north. In 1756, a very strong event (M~7.0) is estimated that has occurred around the area, however there is uncertainty regarding its epicentral locality (Papazachos and Papazachou, 1997). This fault is an obvious candidate to host such a large event due to its large length. Following the above as well as the possible 400 A.D. earthquake (see Noller et al. 1997, but no record in Papazachos and Papazachou 1997) there are at least 4 candidate earthquakes that could have impacted on the Kechriaie subsidence.

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