The uplifted terraces of the Arkitsa region, NW Evoikos Gulf, Greece: a result of combined tectonic and volcanic processes?

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ABSTRACT

The Arkitsa-Kamena Vourla area of central Greece occupies a zone of accommodation between the two tectonic provinces of the North Aegean Trough (the extension of the North Anatolian fault system) and the Gulf of Corinth, and is characterised by a series of very prominent tectonic landforms, notably the large (ca. 1000 m elevation) footwall ridge of the Arkitsa-Kamena Vourla fault system. Despite the highly prominent nature of this footwall ridge and the presence of very fresh tectonic landforms this fault system is not known to have hosted any major historical earthquakes, and the tectonic and geomorphic evolution of the Arkitsa-Kamena Vourla area remains poorly constrained. This paper utilises a combined geomorphological, sedimentological and macro-/micro-fossil approach to evaluate the Late Quaternary evolution of the Arkitsa area, in the eastern part of the fault system, focussing on prominent uplifted terraces present in the hangingwall of the Arkitsa fault. Three distinct raised glacio-lacustrine terraces, and previously reported uplifted marginal marine deposits, suggest sustained uplift of the coastline at a rate of 1 – 1.5 mm/y over at least the last 40,000 years, possibly to 75,000 BP. While movement on an offshore normal fault strand may explain more recent coastal uplift, purely fault-driven longer-term uplift at this rate requires anomalously high fault slip and extension rates. Consequently, the development of the terraces and other geomorphic indicators of uplift may be at least partly due to non-faulting processes, such as Quaternary (intrusive and/or extrusive) volcanic activity associated with evolution of the nearby Lichades volcanic centre.

Keywords: coastal uplift; central Greece; tectonic geomorphology; palaeo-terraces; normal faulting; sea-level
Introduction.

The active normal faulting region of central Greece has been the focus of intense research, due to its relatively high rates of tectonic deformation and the frequent occurrence of damaging moderate magnitude (Ms ≈ 6-7) earthquakes. The structure of central Greece is dominated by a series of roughly WNW-ESE-trending extensional faults. These have created a series of half- (asymmetric) grabens bordered by discontinuous normal faults, the most prominent of which are the Gulf of Corinth and the Evoikos Gulf. Of these two structures, the Evoikos Gulf, and particularly its northern part, is relatively poorly understood in terms of its geodynamic structure and tectonic significance (e.g. Makris et al., 2001). The Arkitsa-Kamena Vourla area occupies the southern part of the northern Evoikos Gulf (Figure 1) and is characterised by a series of very prominent tectonic landforms, notably the large (ca. 1000 m elevation) footwall ridge of the Kamena Vourla fault system. This ridge is one of the major geomorphological features of the northern Evoikos Gulf, and is the surface expression of three major left-stepping (normal) fault segments: the Kamena Vourla, the Agios Konstantinos, and the Arkitsa faults (Roberts and Jackson, 1991; Ganas, 1997; Kranis, 1999) (Figure 2 and 3). While it is clear that the northern Evoikos Gulf occupies a zone of accommodation between the two tectonic provinces of the North Aegean Trough (the extension of the North Anatolian fault system) and the Gulf of Corinth (Mitsakaki et al., 2013; see Papanikolaou and Royden, 2007 for regional summary), the interaction between these provinces is not well understood. Recent GPS measurements indicate that the northern Evoikos Gulf is subject to a relatively low magnitude extensional strain field (Hollenstein et al., 2008) which, apparently contradicting the regional geomorphic evidence, should prohibit the development of large faults. Recent research indicates that seismic stress in this area may not necessarily be released with strong earthquakes, but instead with intense microearthquake activity, usually in swarms (Papanastassiou et al., 2001; Papouli et al., 2006). Indeed, despite the highly prominent nature of the footwall ridge and the presence of very fresh tectonic landforms, the Arkitsa-Kamena Vourla fault system is not known to have hosted any major historical earthquakes (Roberts and Jackson,
and geomorphic) evolution of the wider Arkitsa-Kamena Vourla area remains poorly constrained.

Goldsworthy and Jackson (2001) argue that the Kamena Vourla fault zone is probably a relatively young system, initiated < 1 Ma ago at the expense of the inland Kalidromon fault (located SW of Mt. Knimis, Figure 1 - see Figure 5a of Goldsworthy and Jackson, 2001). Jackson and McKenzie (1999) identify up to 50 increments of coseismic slip on the Arkitsa fault strand, implying that the Arkitsa segment is the active fault trace in the eastern part of the Kamena Vourla fault zone (Dewez, 2003). More recently however Cundy et al., (2010) have identified and described a prominent raised marginal marine unit (most likely a raised beach) and possibly uplifted beachrock in the coastal zone occupying the hanging wall of the Arkitsa fault, near the archaeological site of Alope, which these authors argue indicates recent uplift (in a series of coseismic events) along a possible active offshore fault strand over the late Holocene. Indeed, offshore geological and geophysical surveys carried out in the northern Evoikos Gulf under the recent AMFITRITI project of the Hellenic Centre for Marine Research (HCMR) (http://amphitriti.ath.hcmr.gr) identify several sites of submarine faulting and seafloor ruptures and the presence of an E–W trending normal fault system offshore of Alope / Arkitsa (Figure 2) is inferred. The activation of this inferred offshore fault offers a mechanism to generate the uplift observed by Cundy et al., (2010) along the Alope coast, although seismic and sub-bottom profiling across the north Evoikos Gulf have (as of yet) failed to discriminate any clear evidence for movement on this offshore fault strand (Sakellariou et al., 2007).

Based on radiocarbon dating of preserved marine fauna and numismatic dating Cundy et al. (2010) propose an average coastal uplift rate at Alope over the last 3000 years in excess of 1 mm/y. While it was unclear to what extent this rapid uplift rate had been sustained over the Late Quaternary, the same authors also note the presence of a series of geomorphological features (terraces (previously noted by Mercier, Dewez and others (Dewez 2003)), cones etc.) between the coast and the Arkitsa fault strand which may indicate longer-term uplift. Here, we present the first detailed multi-proxy study of these geomorphic indicators (via a combined geomorphological, sedimentological
and macro-/micro-fossil approach), focussing in particular on three prominent terraces in the hangingwall of the Arkitsa fault, and evaluate the Late Quaternary evolution and uplift dynamics of this important zone of tectonic strain transfer.

Materials and Methods.

Geomorphological mapping.
Detailed geomorphological mapping at a scale of 1:5000 was performed in a series of field seasons between 2007 and 2010, focusing on the prominent terraces around Arkitsa previously noted by Dewez (2003) and Cundy et al., (2010), and other potential tectonic geomorphic features such as knickpoints, incised alluvial cones or fans, etc. In addition, 1:5000 scale topographic maps were digitised and analysed using GIS technology and terraces and other features visualised and delineated using a digital elevation model (DEM).

Stratigraphy.
Exposed sedimentary sections of terrace material (particularly the “cap-rock” of the terrace surfaces) were cleaned and logged, and intact shell material sampled for $^{14}$C dating. The elevations of the exposed units were determined using a Jena 020A theodolite unit, or from 1:5000 scale maps.

Radiometric dating.
Intact carbonate shell material was dated via accelerator mass spectrometry $^{14}$C assay (at the Beta Analytic Radiocarbon Dating Laboratory, Florida, U.S.A.). Radiocarbon age calibration (where possible, i.e. on lower elevation, more recent material) was performed using the MARINE04 database (Hughen et al., 2004), via the programme CALIB 5.0 (Stuiver and Reimer, 1993). A $\Delta R$ value of - 80 ± 25 years was used, corresponding to the local reservoir age correction for Mediterranean surface waters (Stiros et al., 1992; Pirazzoli et al., 1999).

Microfossil analysis.
Approximately 250 g of material from each terrace was processed for microfossil analysis using a variation of the Glauber’s Salt method (Franke, 1922; Wicher, 1942). Indurated sediment samples were soaked in a hot, super-saturated solution of hydrated sodium sulphate and slowly allowed to cool before being frozen overnight. The samples were subsequently defrosted in a microwave oven and the soak-freeze-thaw cycle repeated until the sediments had been sufficiently disaggregated; they were then wet-sieved at >63 μm to remove any finer material. The residue was dried overnight in an oven at 105°C before being examined under a reflected-light binocular microscope (up to 80 x magnification). Specimens were individually hand-picked from small (~5 g) residue sub-samples and mounted on cavity slides. Identification of the ostracod fauna was made with reference to Meisch (2000).

Results.

Geomorphology: Fluvial systems.

There are four important drainage systems (torrents) in the study area (Figure 3). The longitudinal profiles of the four torrents differ significantly as their evolution depends heavily on the fault tectonism of the area (Figure 4). The Alope and Kounoupitsa torrents are located west of Arkitsa and cross the Arkitsa fault, resulting in convex stream profiles. The Alope torrent flows in a SSW-NNE direction and the Kounoupitsa torrent initially has a W-E direction and then deflects towards the North. Both of these torrents perpendicularly cross the Arkitsa fault zone and exhibit prominent knickpoints. The Alope and Kounoupitsa torrents have both heavily incised into Plio-Pleistocene deposits but the Kounoupitsa torrent exhibits an asymmetric transverse valley cross section with steep cliffs on the south side (Figure 4, inset) due to the tilting of the Arkitsa fault footwall and differential erosion. The Kounoupitsa torrent is further characterized by two distinct cones, the older one being of (presumed) Late Pleistocene age located just north of the Arkitsa fault scarp and the second most recent one of Holocene age located near the shore further north. The older cone is deeply incised up to 26 m at the apex of the fan. The reactivation of the Arkitsa fault seems to be almost continuous until the present day (in accordance with Jackson and
MacKenzie 1999) because the almost 100 m high nickpoint of the Kounoupitsa torrent has not moved upstream from the fault scarp. The other two torrents, the Kynos and Livanates torrents (located south of Arkitsa and flowing to the east), are located wholly on the footwall of the Arkitsa fault, and do not exhibit prominent nickpoints. These two torrents have formed normal alluvial cones which are currently eroding. The Kynos and Livanates torrents depict normal concave longitudinal profiles having evolved mostly on easily-erodible Neogene formations (marls, sandstones and conglomerates).

**Geomorphology: Palaeosurfaces and terraces.**

In the footwall of the Arkitsa fault four (presumed) Pleistocene-age palaeosurfaces can be identified (Figure 3) which have evolved on Neogene formations composed of conglomerates, sandstones and marls. The oldest of these extends from 400-480 m amsl (above mean sea level), the second 300-390 m, the third 200-280 m, and the most recent one from 140-170 m. The general dip of these palaeosurfaces is towards the south indicating the continuous activity of the Arkitsa fault system.

The geomorphology in the hangingwall of the Arkitsa fault is dominated by a series of deeply-incised colluvial fans and a number of distinct terraces, at elevations of 2-20 m (terrace A), 25-55 m (terrace B) and 50-80 m (terrace C) amsl (Plate 1, Figures 3 and 5). The formation and evolution of the older terrace (C) is contemporaneous with the Kounoupitsa older cone. The higher terraces in the extreme south of this area, around a proposed secondary fault strand at Livanates, are described by Dewez (2003), with the 25-55 m terrace (terrace B) and 50-80 m terrace (terrace C) (Figure 3 and 5) described in the present study correlating with Dewez’s T1 and T2 terraces respectively. The terraces are cut into Plio-Pleistocene marls, which are unconformably overlain (most clearly in terraces B and C) by well–cemented arenaceous sediments, forming a prominent “cap–rock”. Detailed descriptions of this cap-rock are given below. The two higher and older terraces are deeply dissected down to 27 m by recent fluvial activity. In addition, north of the Arkitsa fault scarp (at 38° 43.680’ N; 22°59.340’ E) a series of small talus cones have formed during late Pleistocene-Holocene times, which are also deeply incised (Figure 3).
Terrace stratigraphy: Terrace A

This terrace has developed east of the village of Arkitsa and extends for more than 3 km in a NW-SE direction. The terrace has a clear dip to the NE of ~2 %. Its inner edge is found at an elevation of about 20 m. A prominent exposure of terrace A was located at 38° 45.128’ N; 23° 01.753’ E, at +6 m elevation. While the section was partly-obscured and overgrown, the exposure consisted of 90 cm of highly fossiliferous coarse sandstone unconformably overlying 17 cm of Plio-Pleistocene white silty-clay. The silty-clay unit (in the lower part of the exposure) was relatively unconsolidated, and contained no apparent bedding. The arenaceous cap-rock in the upper part of the exposure consisted of a coarse, well-rounded and sorted sand with grit and occasional small (<1 cm) sub-rounded limestone pebbles. Macrofossil remains were common and included both intact and disaggregated bivalve and gastropod shell material, with identifiable fragments (ca. 5mm in size) consisting entirely of juvenile Mytilus sp. (possibly Mytilus galloprovincialis). Shell remains were distributed highly heterogeneously; in some parts they were almost absent, whereas in other areas they comprised 30–50 % of the cap-rock. Occasional disarticulated juvenile ostracod remains of Cyprideis torosa were present, as well as rare single valves of juvenile Candona sp.

Terrace stratigraphy: Terrace B

This terrace is the most prominent in the area, and extends for 7.5 km in an E-W direction. Its elevation increases towards the east, so the inner edge rises from a height of 25 m in the west to 55m in the east. The terrace is tilted towards the south with a dip of 1 %. A clear roadside exposure of terrace B was found at 38° 44.517’ N; 23° 01.823’ E, at +30 m elevation. The exposure consisted of a 1 m-thick unit of well-cemented sandstone, unconformably overlying a lower unit (1.6 m thick) of thinly-bedded marly silts and sands. Bed thickness in this thinly-bedded lower (Plio-Pleistocene) unit was 1 – 5 cm, dipping at 16° WNW. The upper sandstone (cap-rock) unit consisted of a well-cemented medium sand and silt, with sub-rounded to rounded grits and pebbles of limestone and chert composition. Pebbles ranged between 1 and 3 cm in diameter. Small (< 1 cm) isolated gastropod fragments (apparently Viviparus sp.)
were locally present. A more complete and extensive sequence of the well-cemented arenaceous cap-rock was found in a roadside cut and adjacent private land at 38° 44.589’ N; 23° 02.002’ E, showing 1.8 m of exposure, consisting of 0.55 m of well-cemented sandstone (stratigraphy as detailed above) overlying 1.25 m of bedded *Viviparus* sp.-rich well-cemented medium to coarse sand, with sub-angular to sub-rounded chert (and occasional limestone) pebbles, up to 3 cm diameter. Intact *Viviparus* sp. shells occurred throughout this unit, locally concentrated into coquina-type beds. Juvenile valves of the ostracods *Cyprideis torosa* and *Candona* spp. were common throughout, valves of *Ilyocpris* sp. and *Tyrhenocythere* sp. (also juvenile) were also present but were considerably rarer.

**Terrace stratigraphy: Terrace C**

This terrace extends for 5 km in an E-W direction. Its elevation increases towards the east, so the inner edge from a height of 50 m in the west reaches 80 m in the east. The terrace is tilted towards the south having a dip of 1.5 %. A prominent exposure in a road cut (adjacent to the main access road to Arkitsa port) at 38° 44.622’ N; 23° 01.567’ E, +50 m elevation, showed ca. 0.5–0.7 m of well-cemented, gastropod-rich, sandstone unconformably overlying yellow-white marls. The stratigraphy of the lower (Plio-Pleistocene) marls is described in detail in Dewez (2003). The upper sandstone (cap-rock) unit was rich in intact and fragmented *Viviparus* sp. (Dewez’s “*Viviparus cayrock*”) and other gastropods (including rare *Valvata piscinalis*), sometimes forming coquina-type lenses, in a medium to coarse sand matrix. Ostracod remains were rare, however, and limited to occasional juvenile, disarticulated valves of *Cyprideis torosa* and *Candona* spp.. The sand contained sub-rounded to rounded limestone and chert grit clasts, and rounded to sub-rounded limestone, chert and (sporadic) mafic pebbles, of 4–5 cm diameter. The structure of the cap-rock layer here was relatively complex, as the ca. 100 m of horizontal exposure indicated the presence of two unconformable cap-rock layers (Plate 2), an upper (horizontally-bedded) unit above a lower (slightly older) cap-rock unit with bedding dipping at 5° E. At a small quarry N of this road cut (38° 44.642’ N; 23° 01.528’ E), where the cap-rock has been removed for use locally as a building stone, intact *Viviparus* sp. fragments were less frequent, although the section contained a series of highly abraded broken shell or coquina layers.
Radiocarbon dating

Two valves of *Mytilus galloprovincialis* collected from terrace A (at +6 m) gave conventional radiocarbon ages of 36,200 ± 340 BP and 40,530 ± 500 BP (Table 1). An intact *Viviparus* sp. specimen from terrace B (at +30 m) gave a further conventional radiocarbon age of >43,500 BP (Table 1).

Discussion

The main factor controlling the late Quaternary deposits of the Northern Evoikos Gulf coast and sea floor was global sea level change caused by the numerous glacial-interglacial cycles of the Pleistocene. Exchange of water masses with the Aegean Sea is restricted via (a) the narrow and shallow Oreos-Trikeri straits, which have a maximum depth of approximately 45 m and a mean width of 4 km, and (b) the artificially maintained Euripus Channel (length 60 m, width 40 m, depth 8.5 m) in the southeast of the region (Figure 1). The maximum pre-Holocene sill depth in the Oreos-Trikeri straits is currently approximately 45 m below sea-level (bsl), although Van Andel and Perissoratis (2006) argue for a late glacial sill depth at about 55 m bsl based on high-resolution seismic profiling data. The Northern Evoikos Gulf was isolated from the Aegean Sea at ca. 50,000 yrs BP (Figure 6) and became a lake during Oxygen Isotope Stage (OIS) 3 and 2. The lake began to fill with seawater as soon as sea level reached the critical sill depth at the Oreos-Trikeri straits at about 10,500 $^{14}C$ BP (Lambeck, 1996).

Of the terraces described here, only the lowest (terrace A) is datable via $^{14}C$ (Table 1); the higher terrace B gives a date which exceeds the dating range of $^{14}C$ (> 43,500 BP). While further dating using U-series methods may better constrain the ages of the upper terraces B and C, given the thin, delicate nature of the preserved *Viviparus* fossils, and the high likelihood that these gastropods may act as open systems in terms of U-exchange, such dating is unlikely to definitively ascribe an accurate age to the upper cap rocks. The $^{14}C$ dates from terrace A (ca. 36,000 – 41,000 BP) specify its formation at a period of low regional and global sea-level, on the margins of the
Evoikos glacial lake. While water levels in the glacial lake clearly fluctuated due to climatic variations, the dates for terrace A correlate with OIS-3, a period with warmer and wetter intervals, which Van Andel and Perissoratis (2006) tentatively correlate with a cluster of erosional submarine terraces incised in the Evoikos Gulf floor between 50 and 70 m bsl. The sedimentology of the terrace A cap-rock (i.e. coarse, well-rounded and sorted sand with grit and occasional small [<1 cm] sub-rounded limestone pebbles) indicates deposition in the higher-energy lake shallow margins, although the dominant faunal remains, those of juvenile *Mytilus* sp. (possibly *Mytilus galloprovincialis*), indicate deposition in at least a brackish environment (e.g. Ceccherelli and Rossi 1984). The age of this terrace correlates with a period of relative high stand, when sea-levels were close to the sill depth (Shackleton 1987, Labeyrie et al 1987 and discussion in Van Andel and Perissoratis 2006), which may explain the dominance of *Mytilus* sp.. This interpretation is further supported by the microfossil assemblage. *Cyprideis torosa* are euryhaline ostracods, tolerating a broad range of salinities from almost freshwater to hypersaline (but they are never fully marine) (Meisch, 2000). They tend to occupy estuaries, lagoons, brackish-water bays and shallow near-shore littoral environments that have sandy/muddy substrates with some algal/organic cover. Their valves are moderately robust and so they can tolerate moving water and moderate energy conditions, further evidenced here by a lack of larger instars. On the other hand, candoniid ostracods are not usually associated either with elevated salinities or high-energy environments (Meisch, 2000), suggesting that the rare *Candonia* sp. valves here may have been washed in from the local catchments.

Regardless of the precise (palaeo) lake level, the current elevation of terrace A compared to the estimated late glacial Oreos-Trikeri sill depth of 55 m bsl indicates considerable uplift over the last ca. 40,000 years. Based on the age of the *Mytilus* sp. at +6 m and a sill depth of 55 m bsl, a minimum average uplift rate of 1.5 mm/y can be derived (i.e. 61 m in 40,000 years, which is a minimum estimate due to uncertainties in palaeo-lake level). Considering the higher terraces, while the ages of terraces B and C are not well-constrained (> 43,500 BP), their sedimentological characteristics (arenaceous to conglomeratic) and fossil content indicates their formation in a
freshwater lake margin. *Viviparus* sp. prefer shallow freshwater or (occasionally) oligosaline habitats with muddy substrates such as might be found on the margins of slow-moving lowland rivers and the littoral areas of large lakes (e.g. Glöer, 2002).

Coupled with the presence of freshwater (*Candona* spp., *Ilyocypris* sp.) and oligosaline (*Cyprideis torosa*, *Tyrrenocythere* sp.) ostracods (e.g. Meisch, 2000; Griffiths, 2002; Alvarez-Zarikian, 2008), the faunal evidence points to a shallow, freshwater or very slightly brackish lake / lagoonal margin. Based on the discussion above, it is likely that these terraces were formed in earlier phases of OIS 3 and OIS 4 during relative high stand periods, and then uplifted to their present elevations. Indeed, Van Andel and Perissoratis (2006) note a series of submerged erosional terraces in the northern Evoikos Gulf clustered around discrete depth intervals, and probably correlating with lake level fluctuations over OIS 2-4. Sea-level curves (Figure 6) suggest that terrace B may have been formed around 55,000–60,000 yrs BP and terrace C prior to about 70,000 yrs BP. Further support for this inference comes from the terrestrial pollen record, including evidence from the well-dated lacustrine sequences from Ioannina in the north-west of Greece (Tzedakis et al., 2002) and Kopais on the Boetian plain, located less than 40 km south of Arkitsa (Tzedakis, 1999). Palynological data from both of these sites indicate the presence of intermediate forest (including a temperate tree component) for part of the Middle Pleniglacial between *ca.* 50,000 and 59,000 yrs BP (encompassing the proposed formation of terrace B), and fully-forested interstadial conditions between *ca.* 68,000 and 83,000 yrs BP (encompassing the proposed formation of terrace C). During these relatively wet and warm periods, sea-level was at a depth of *ca.* 60 m below present sea level or higher, below but not far from the proposed sill depth of 55 m. If correct these estimates would suggest that terrace B was uplifted by *ca.* 85 m (i.e. 60 m + 25 m [lowest position of the inner terrace edge]), and terrace C by *ca.* 110 m (i.e. 60 m +50m [lowest position of the inner terrace edge]), having uplift rates of about 1.4 mm/y (85 m/60,000 yrs for terrace B and 110 m/75,000 yrs for terrace C).

Terrace data therefore suggest uplift at average rates of 1–1.5 mm/y beyond the Holocene in the late Pleistocene, to at least 40,000 yrs BP. This estimated uplift rate is significantly higher than the long-term (footwall) uplift rate of 0.2 mm/y calculated by
Goldsworthy and Jackson (2001) for the eastern end of the Arkitsa fault segment (also reported by Walker et al., (2010)), and while it agrees well with Late Holocene uplift rates of 1–1.4 mm/y proposed by Cundy et al., (2010) based on nearby slightly uplifted marginal marine units around Alope, it poses significant problems in terms of fault dynamics and extension rates in the area. While rupture of the offshore fault inferred by Sakellariou et al., (2007) and Cundy et al., (2010) provides a mechanism for recent coseismic coastal uplift in the Alope area, sustained coastal uplift on this fault of 1–1.5 mm/y requires anomalously high slip and extension rates. Specifically, if a footwall-hangingwall partition of 1:2 is adopted (following McNeill et al.’s (2005) range of 1:1.2 to 1:2.2 derived from active faults in the Gulf of Corinth rift, which have similar dip to those in the Arkitsa area), then the resultant slip rate on this offshore fault would be ca. 4 – 6 mm/y, with an extension rate of 3 – 4.5 mm/y. This slip rate is much higher than recent estimates using continuous GPS stations in central Greece (Chousianitis et al., 2013), which (a) indicate that most of the extensional strain across the northern Evoikos Gulf is accommodated by the Arkitsa fault, and (b) suggest at most a 1.2–1.5 mm/y slip rate on other active faults in this area. The high uplift rate estimates are similarly problematic given that renewed activity on the onshore Arkitsa fault will down-throw the terraces, and require even higher slip rates on offshore fault systems.

In this respect, of note is the apparent presence in terrace C of two unconformable cap-rock layers (Plate 2) with differing bedding dip angles. The contrast in dip angle is laterally extensive and, although the section is partly obscured by vegetation, the unconformable relationship between the two units is clear (Plate 2). This indicates formation of the cap-rock here in at least two phases, interrupted by an episode of slight tilting and erosion, possibly related to local coseismic movement from the activity of the Arkitsa fault. In other words, after the formation of the lower cap-rock, it is possible that the Arkitsa fault reactivated and caused the tilting of this formation towards the fault. The proximity of this site (some tens of metres) to the scarp of the Arkitsa fault and its position on the fault’s hanging wall supports this hypothesis. Subsequently, the newer cap-rock was formed in discordance with the former one. This event should have happened during the formation of terrace C, i.e. possibly around 70,000–75,000 yrs BP. The existence of a fresh strip of fault scarp at the base of the Arkitsa fault, attributed to Holocene activity (Jackson and MacKenzie 1999), and
the geomorphology of the fluvial systems which cross the fault, indicates that the
Arkitsa fault remains active.

Given the excessively high, long-term offshore fault slip rates required to generate
purely extensional fault-driven uplift of the terraces to their present elevation, some
other explanation must therefore be found for their presence at up to 80m above
current sea-level. Possible causes are non-extensional local uplift, or formation of the
terraces at elevations above the contemporary shoreline due to local damming of the
Evoikos glacial lake. Two volcanic centres are present (the Lichades and Chronia
volcanic centres) to the northwest and northeast of the Arkitsa area which were
activated during the Pliocene and Quaternary (Pe-Piper and Piper 1989, 2002;
Lambrakis and Kalergis, 2005, although the Chronia volcanic centre is of questionable,
possibly older, age, [Karasththis et al., 2011]), and a magma chamber at 8km depth has
been detected north of Arkitsa (Karasththis et al., 2011) which is argued to be the
source of the widespread hydrothermal activity observed in the northern part of the
Evoikos Gulf. Basaltic sill intrusion could potentially generate local uplift, and extrusion
of lava cause local shoreline changes, particularly near to the Oreos-Trikeri straits, and
“barriering” or damming of lakes at higher elevations. The exact causal mechanism of
terrace development and uplift remains enigmatic, and further detailed studies of
these volcanic centres and their influence on palaeo-shoreline geomorphology and
local uplift (coupled with offshore geophysical work in the NW Evoikos Gulf to better
constrain offshore fault slip and activity and the relationship between onshore and
offshore (submerged) terraces) are needed to better constrain the Late Quaternary
uplift pattern in this area, and determine the relative roles of extensional faulting and
other regional processes (e.g. volcanic centre activity) in driving palaeoterrace
formation and uplift.

Conclusions

A combined geomorphological, sedimentological and macro-/micro-fossil approach
has been used to evaluate the Late Quaternary tectonic evolution and uplift of the
Arkitsa area, in the eastern part of the Arkitsa-Kamena Vourla fault system, focusing on
prominent terraces present in the hangingwall of the Arkitsa fault. Three distinct raised
glacio-lacustrine terraces, and previously reported uplifted marginal marine deposits, indicate sustained uplift of the coastline at a rate of 1 – 1.5 mm/y over at least the last 40,000 years, possibly to 75,000 BP. While movement on an offshore normal fault strand may partly explain this (and more recent) coastal uplift, purely fault-driven uplift requires anomalously high fault slip and extension rates over the Late Quaternary. Consequently, the development of the terraces and other geomorphic indicators of uplift may be at least partly due to non-extensional faulting mechanisms, such as Quaternary (intrusive and/or extrusive) volcanic activity associated with evolution of nearby volcanic centres. Further detailed studies of these volcanic centres and their influence on palaeo-shoreline geomorphology and local uplift, coupled with offshore geophysical work in the NW Evoikos Gulf to better constrain offshore fault slip and activity, are needed to determine the relative roles of extensional faulting and other processes in driving palaeo-terrace formation and uplift around Arkitsa.

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References cited.


Figures and Tables:

Table 1: Radiocarbon and numismatic dates from the Arkitsa terraces and uplifted marginal marine units exposed around Alope.

Figure 1. Location of the Northern Evoikos gulf in central Greece. Relief shading shows the general bathymetry of the Gulf and the topography of the coast and inland areas. The location of the Oreos-Trikeri straits in the north and Euripus channel in the south, and location and names of major fault systems (K.V.: Kamena Vourla, A.G.: Agios Konstantinos, Ar: Arkitsa, At.: Atalanti, T: Tragana, M: Malesina, Ef.: Evia fault zone) are also marked. Stars indicate the Lichades (L) and Chronia (C) volcanic centers. Projection/datum used: Hellenic Geodetic Reference System 1987 - non-geocentric datum.

Figure 2. Geological setting and tectonic structure of the Arkitsa-Kamena Vourla area, north Evoikos Gulf, central Greece. After Institute of Geology and Mineral Exploration (IGME) sheets (1:50,000 scale): Elateia (1967), Pelasgia (1957), Livanatai (1961) and Istiaia (1984).

Figure 3. Geomorphological map of study area: Kynos-Arkitsa-Alope coastal zone. Topographic data taken from 1:5000 scale topographic maps (Hellenic Military Geographical Service, 1979). Line CS_1 shows location of cross-section for Figure 4 inset.

Figure 4. The longitudinal profiles of the Alope, Kounoupitsa, Kynos and Livanates torrents. Inset gives the N-S cross section of the Kounoupitsa torrent, marked as line CS_1 in Figure 3.

Figure 5: Digital elevation model (20m x 20m) of the Arkitsa area, highlighting terrace A (green), B (orange) and C (blue). Vertical exaggeration is x5. Google Earth imagery reference: Google earth V 7.1.2.2041. (10/7/2013). Arkitsa, Greece. 38°44.041’ N; 22°59.735’ E; Eye alt. 1500m. Digital Globe 2014.

Figure 6. Global eustatic sea-level curves for the last ca. 80 kyr after Shackleton (1987) and Waelbroeck et al., (2002), illustrating major global sea-level trends during the Late Quaternary. OIS 3 and OIS 4 are labeled, and the estimated depth of the Oreos-Trikeri sill during late glacial periods is also marked (after Van Andel and Perissoratis 2006).

Plate 1: Panoramic view of the 3 prominent terraces in the area of Arkitsa (view east, from 38°44.50’ N; 23°00.50’ E). From left to right: terrace A (2-20 m), terrace B (25-55 m), terrace C (50-80 m).

Plate 2: Panorama of terrace C exposure (view towards east, from 38° 44.622’ N; 23° 01.567’ E), showing two unconformable layers of cap-rock. Note contrast in dip angle (see text for further discussion). Figure for scale is 1.8m in height.
<table>
<thead>
<tr>
<th>Sample</th>
<th>Species / type</th>
<th>Sample elevation (above HWL)</th>
<th>Conventional $^{14}$C age (a BP, ± 1 $\sigma$)</th>
<th>Lab. number</th>
<th>$^{14}$C/$^{12}$C ($\delta^{13}$C)</th>
<th>Calibrated age (a BP)$^2$</th>
<th>Source</th>
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</thead>
<tbody>
<tr>
<td>Shell</td>
<td><em>Viviparus</em> sp.</td>
<td>30m</td>
<td>&gt;43500 BP</td>
<td>Beta-285990</td>
<td>-7.0</td>
<td>-</td>
<td>This study</td>
</tr>
<tr>
<td>Shell</td>
<td><em>Mytilus</em> sp. (prob. <em>M.galloprovincialis</em>)</td>
<td>6m</td>
<td>40530 ± 500 BP</td>
<td>Beta-258442</td>
<td>+0.6</td>
<td>-</td>
<td>This study</td>
</tr>
<tr>
<td>Shell</td>
<td><em>Mytilus</em> sp. (prob. <em>M.galloprovincialis</em>)</td>
<td>6m</td>
<td>36200 ± 340 BP</td>
<td>Beta-285989</td>
<td>-0.5</td>
<td>-</td>
<td>This study</td>
</tr>
<tr>
<td>Shell</td>
<td><em>Cerithium</em> sp.</td>
<td>1.1 – 1.6 m</td>
<td>2900 ± 40 BP</td>
<td>Beta-218939</td>
<td>+0.8</td>
<td>916 – 724 BC</td>
<td>Cundy et al. (2010)</td>
</tr>
<tr>
<td>Shell</td>
<td><em>Cerithium</em> sp.</td>
<td>1.1 – 1.6 m</td>
<td>2980 ± 40 BP</td>
<td>Beta-218940</td>
<td>+0.7</td>
<td>1001 – 783 BC</td>
<td>Cundy et al. (2010)</td>
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<tr>
<td>Shell</td>
<td><em>Spondylus</em> sp.</td>
<td>0.9 – 1.5 m</td>
<td>3430 ± 40 BP</td>
<td>Beta-236948</td>
<td>+1.6</td>
<td>1581 – 1332 BC</td>
<td>Cundy et al. (2010)</td>
</tr>
<tr>
<td>Coin</td>
<td>Base metal Roman <em>nummus</em></td>
<td>1.1 – 1.6 m</td>
<td>–</td>
<td>–</td>
<td>–</td>
<td>AD 378 – 383$^3$</td>
<td>Cundy et al. (2010)</td>
</tr>
<tr>
<td>Shell</td>
<td>Gastropod (sp. indet.)</td>
<td>0.30 m</td>
<td>1820 ± 40 BP</td>
<td>Beta-236949</td>
<td>-0.1</td>
<td>AD 394 - 623</td>
<td>Cundy et al. (2010)</td>
</tr>
</tbody>
</table>

$^1$ Shell samples pretreated by etching with HCl.

$^2$ Calibration performed using the MARINE04 database (Hughen et al 2004), using the programme CALIB 5.0 (Stuiver and Reimer, 1993). A delta R value of -80 ± 25 years was used, corresponding to the local age reservoir of Mediterranean surface waters (Stiros al et al 1992, Pirazzoli et al 1999). Values presented show a 2$\sigma$ error margin.


Table 1
Figure 2
Figure 3
Figure 6

Depth (m)

present sea level

Waelbroeck 2002

Sill level

Shackleton 1987

Age (kyr)