Testing rival tectonic uplift models for the Lechaion Gulf in the Gulf of Corinth rift

J. A. TURNER1*, M. R. LEEDER1, J. E. ANDREWS1, P. J. ROWE1, P. VAN CALSTEREN2 & L. THOMAS2

1School of Environmental Sciences, University of East Anglia, Norwich NR4 7TJ, UK
2Open University Uranium Series Facility, Department of Earth Sciences, The Open University, Walton Hall, Milton Keynes MK7 6AA, UK

*Corresponding author (e-mail: jenni.turner@uea.ac.uk)

Abstract: The Gulf of Corinth, central Greece, is a rapidly extending continental rift, the eastern part of which bifurcates into the active northern Alkyonides Gulf and the southern Lechaion Gulf. The Lechaion Gulf is considered an inactive relict of early rifting, yet the presence of late Quaternary shorelines is evidence of continuing uplift of the north, east and south margins of this basin. Models to explain uplift include uplift on the footwall of the southern Alkyonides Gulf fault system and the Xyloastro–Perachora faults or ‘regional’ isostatic uplift independent of fault slip. These models are tested by comparing predicted spatial uplift trends with those observed. Uplift rates since Marine Isotope Stage 7 of 0.31 ± 0.04 mm a−1 on the Lechaion Gulf north coast are explained as displacement on the footwall of active faults. However, the south coast uplift cannot be explained by footwall uplift and is evidence for isostatic uplift that probably affects the whole of the southern Gulf of Corinth rift. Isostatic uplift rates of 0.22 ± 0.01 mm a−1 at the Corinth canal increase westward towards the mouth of the Lechaion Gulf where it meets the modern rift, the Gulf of Corinth.

Supplementary material: Details of Holocene marine notches that identify raised shorelines from exposures on the southern Perachora peninsula between Heraion and Loutraki, and of the cumulative displacement of sequences by minor faults on the Isthmus, exposed in the Corinth canal walls, are available at http://www.geolsoc.org.uk/SUP18429.

Uplift of continental and proto-oceanic rift flanks is a common phenomenon worldwide; for example, in Ethiopia–East Africa and the Suez–Red Sea (e.g. Ebinger et al. 1999; Daradich et al. 2003). However, hypotheses to explain such uplift are often difficult to test in the absence of spatial and temporal information on uplift rates. Here we use raised, dated, late Pleistocene and Holocene marine shorelines over a large (c. 350 km2) area of the Lechaion Gulf, a sub-basin in the eastern Gulf of Corinth rift, Central Greece to test previously published uplift hypotheses. The Gulf of Corinth rift is a much studied example of rifting in an overall subduction-zone setting (Fig. 1). Extensional stresses are variously attributed to strain caused by upper plate differential velocity of the Anatolian–Aegea plate (e.g. Le Pichon & Angelier 1979; Billiris et al. 1991; McClusky et al. 2000), back-arc extension from subduction of the African plate beneath Aegea (Vita-Finzi 1993), collapse of a precursor orogen (Meijer & Wortel 1996; Hatzfeld et al. 2003). Diffuse crustal thinning of Aegea probably began in the Miocene with subsequent focused neotectonic extension at a rapid rate of c. 10 mm a−1 in the Gulf of Corinth (Leeder et al. 2003). Diffuse crustal thinning of Aegea probably began in the Miocene with subsequent focused neotectonic extension at a rapid rate of c. 10 mm a−1 in the Gulf of Corinth (Clarke et al. 1997; Brioie et al. 2000; McClusky et al. 2000).

The Corinth rift, trending WNW–ESE, is connected to the Ionian sea in the west, and at the east end bifurcates into two sub-basins, the Alkyonides Gulf and the Lechaion Gulf (Fig. 1b); we use the term ‘Gulf of Corinth’ to refer to the rift that lies west of this bifurcation. The two sub-basins are separated by the Perachora peninsula. Extensional strain is taken up on the faults to the south of the Gulf of Corinth and Alkyonides Gulf. However, the Lechaion Gulf with its exhumed margins (Fig. 1c), the subject of the present paper, is not considered to be part of the active rift but of a proto-Gulf of Corinth (Sakellariou et al. 2004; Leeder et al. 2005). Basinward migration of north-dipping fault systems has resulted in flights of marine terraces and shoreline features from >800 m elevation down to the modern coastline on the southern rift flank of the Gulf of Corinth and Alkyonides Gulf (e.g. Armijo et al. 1996; McNeill & Collier 2004). Late Pleistocene to Holocene rates of rift flank uplift vary from c. 0.8 mm a−1 in the west to 2.0 mm a−1 in the centre and c. 0.3 mm a−1 in the east. Emergent fossil shorelines and their deposits of Late Pleistocene to Holocene age have also been mapped to elevations of c. 400 m around the Lechaion Gulf (Keraudren & Sorel 1987; Collier 1990; Armijo et al. 1996; Leeder et al. 2003, 2005). These are unconformable upon older Pleistocene and Pliocene basin sediments on the south and east margins and on pre-rift Mesozoic basement lithologies of the Perachora peninsula of the northern Lechaion Gulf.

Because the Lechaion Gulf is not considered part of the active rift, palaeoshore and basin-fill emergence has been assumed to be occurring in the absence of extension south of the Perachora peninsula. Various suggestions have been proposed to explain this uplift based on tectonic processes external to the basin (Jackson et al. 1982; Armijo et al. 1996; Morewood & Roberts 1999; Leeder et al. 2003; Leeder & Mack 2008). The present study sets out to test the various hypotheses by comparing
measured uplift trends with those predicted. Measured uplift values are based on published data but where data are contradictory or absent we have mapped evidence for raised shorelines and dated samples to calculate uplift rates. These allow us to determine the spatial distribution of uplift rates for comparison with those predicted by each model. Also, as there is no clear consensus about the activity status of crustal-scale faults bordering the Lechaion Gulf (e.g. Roberts & Jackson 1991; Goldsworthy & Jackson 2001; Moretti et al. 2003; Sakellariou et al. 2004; Leeder et al. 2005) we also use our data to help determine the activity status of the faults and to resolve the apparent discrepancy of active faulting with the assumption noted above that the basin is no longer actively extending.

**Tectonic uplift models**

The first hypothesis for uplift of the Lechaion Gulf was suggested by Jackson et al. (1982; see also Goldsworthy & Jackson 2001). This postulates that footwall uplift occurred across the study area as a result of displacement along the active Pisia–Skinsos fault system (Fig. 2).

A second hypothesis (Armijo et al. 1996; supported by Dia et al. 1997) arose from an elastic-slip model proposed for displacement on the offshore Xylokastro fault, considered to continue to the north of the Perachora peninsula, with uplift in the footwall of this structure, including the Lechaion Gulf, decreasing away from the fault trace (Fig. 2). Marine seismic surveys (Stefatos et al. 2002) have since identified this supposedly continuous fault line as comprising discrete Xylokastro and Perachora faults. Moreover, Armijo et al.’s (1996) fault-slip model yields displacements that far exceed the depth to pre-rift basement of the Gulf of Corinth (Sakellariou et al. 2004; Leeder et al. 2005), so here we adapt their model to reflect known fault configuration whilst retaining the general concept of footwall uplift decaying with distance from the Xylokastro and Perachora faults.

These two models predict uplift associated with specific faults, but as both are located on the southern margin of the actively rifting Alkyonides Gulf, extensional strain is probably taken up across the fault segments in a combined linkage. We thus combine the two models and present a quantitative pseudo-3D model of the combined contribution of these active fault systems to uplift of the Perachora peninsula and Lechaion Gulf (Fig. 3). As the original models lack quantitative data, we calculate values using the average footwall uplift rates for the Perachora peninsula. As uplift rates may vary over the shorter term we apply the rate of 0.3 mm a⁻¹ calculated from the Marine Isotope Stage (MIS) 7 shoreline (Dia et al. 1997; Collier et al. 1998; Leeder et al. 2003) and for scaling purposes model displacement of a 320 ka (MIS 9) shoreline. To estimate hanging-wall subsidence we adopt a footwall:hanging-wall partitioning of 1:2, after McNeill et al.’s (2005) range of 1:1.2 to 2.2, derived for the steeply dipping faults of the Gulf of Corinth rift: this maximizes net uplift values as any competing effects of hanging-wall subsidence are minimized. We assume that footwall uplift decays to zero at a distance from, and normal to, the fault trace that is approximately equal to both the fault depth (Jackson et al. 1982) and local thickness of the seismogenic brittle upper crust, 8–15 km (Jackson et al. 1988; Roberts & Jackson 1991; Taymaz et al. 1991). We apply 15 km decay distance to the c. 15 km long Pisia fault trace but reduce the estimate to 8 km decay distance for the shorter (3 km long) Upper Loutraki fault. The effects of the active Heraion and Upper Loutraki faults that define the southern boundary of the Perachora peninsula (discussed below) are also accounted for. The result is a spatially variable displacement of the topography associated with each fault, which we combine to derive a net value along the three cross-sections of the Lechaion Gulf. This produces a pseudo-3D representation of the hypothetical model predictions (Fig. 3a).

In a departure from a local uplift mechanism, Leeder et al. (2003) proposed spatially uniform ‘regional uplift’ of the Lechaion Gulf and Perachora peninsula, a concept developed from the isostatic uplift suggested by Collier et al. (1992). Leeder et al. (2003) and Leeder & Mack (2008) attributed this ‘regional uplift’ to an isostatic adjustment of the lithosphere.
independent of fault slip. The envisaged mechanism for this regional component of uplift and the Quaternary rifting of the Gulf of Corinth is Aegea being pushed over the hinge of the steeply dipping section of the subducted African plate (Fig. 1c) onto the shallow-dipping section below the Peloponnese. Isostatic uplift, which is uniform across the Lechaion Gulf, is also modelled since MIS 9 using an uplift rate of 0.3 mm a$^{-1}$ (Leeder et al. 2003; Fig. 3b).

These models lead to differing predicted rates and patterns of uplift for the Lechaion Gulf north and south margins (Fig. 3), which can be tested with in-field data (elevations and ages) from raised shorelines.

**Mapping palaeoshorelines: method and results**

Emergent palaeoshorelines have been widely used as reference markers for displacement of coastal landmasses (e.g. Jackson et al. 1982; Lajoie 1986; Gray & Ivanovich 1988; Pirazzoli et al. 1994; Armijo et al. 1996; Leeder et al. 2003; Schellmann et al. 2004). Rigorous quantification of uplift rate requires: (1) a precise modern sea-level reference datum that we measure at modern high water mark with an accuracy of ±0.05 m during calm sea conditions; (2) a well-constrained palaeo sea-level feature, for which we use the mean water mark of a beach profile at the shoreline ‘inner edge’; (3) accurate palaeoshoreline chronology to calculate time over which uplift has occurred. Ages are from thermal ionization mass spectrometry (TIMS) U–Th analyses of the Mediterranean coral *Cladocora caespitosa*, collected from *in situ* coral thickets in sediment sequences that are correlated to mapped inner edges, and $^{14}$C dating of *Lithophaga* shells from marine notches; (4) an adjustment for differences in palaeo sea level of each highstand with modern sea-level datum for which we use the sea-level curve of Thompson & Goldstein (2006). Any glacioisostatic or hydroisostatic influence on palaeoshoreline elevation is expected to affect the Lechaion Gulf uniformly; these lithospheric-scale adjustments are not significant when comparing shoreline elevations across this c. 20 km wide basin.

Elevations measured from the high water mark to 35 m were made by Abney level and, unless otherwise stated, higher elevations were measured using a Barigo altimeter and cross-checked with a global positioning system (GPS) reading and a position fix verified with Hellenic Military 1:5000 topographic maps to ±5 m difference. Tectonic uplift rates were calculated from shoreline elevation and age after the method applied to Barbados coral terraces (e.g. Schellmann et al. 2004; Speed & Cheng 2004). We include uncertainties for calculated uplift rates that include: (1) a measurement error of ±0.1 m for Holocene inner edges elevation and of ±2.0 m for shorelines older than the Holocene, unless otherwise stated; (2) the palaeoshoreline represents the duration of the highstand (MIS 5a ±7 ka, MIS 5c ±2 ka, MIS 5e 4–5 ka, MIS 7a ±10 ka, MIS 7c ±4 ka) which is typically greater than age analysis uncertainty; (3) uncertainty of sea-level elevation. As these sea-level elevation uncertainties are not always shown on sea-level curves, we compared the global sea-level curve with the regional curve compiled by McNeill & Collier (2004). The uncertainties represent the variation of highstand elevation by ±5 m, MIS 5a and 5c ±5 m, MIS 5e ±5 m, MIS 7a ±10 m, MIS 7c ±5 m, and MIS 9 ±10 m.

Detailed palaeoshoreline mapping around the study area from the Perachora peninsula to the Corinth canal was combined with published data that we verified at key locations and refer to in the results and discussion sections. On the Perachora peninsula at 0–4 m elevation, marine notches formed in the intertidal zone are preserved at different elevations in well-lithified rock and cemented raised beach sediments. Limestone substrates frequently yield holes bored by the marine mollusc *Lithophaga*, which typically characterize a linear near-horizontal upper level

---

**Fig. 2.** Map showing locations referred to in the text, fault names, and the uplift models. The fault slip models describe uplift of the Loutraki and Corinth plains on the footwall of active faults (arrows), the Xylokastro and Perachora (Armijo et al. 1996) and the Alkyonides fault system (Pisia and Skinos) (e.g. Jackson et al. 1982). Leeder et al. (2003) proposed uniform isostatic uplift of the region (+).
Fig. 3. Predicted uplift of the Lechaion Gulf since MIS 9 (320 ka). (a) Fault slip uplift; each cross-section depicts scaled footwall and hanging-wall displacement for each fault with the net cumulative displacement from combined uplift and subsidence represented by the dotted line. (b) Uplift modelled by the isostatic uplift model of Leeder et al. (2003).
at the notch recess. Holocene marine platforms several metres wide may be cut in Mesozoic basement limestone, serpentinite, or cemented cobbles and pebbles; some are backed by notches that more accurately define the inner edge. Overall, preservation of shorelines and beach deposits decreases with increasing age, thus only the most robust features are found at higher elevations. The elevation, location and type of shoreline feature mapped are summarized in Figure 4. New U-series ages of *C. caespitosa* are summarized in Table 1.

When calculating regional trends in uplift rates using discrete palaeoshoreline occurrences we also consider the effect of localized minor faulting that displaces stratigraphy by 1 m to tens of metres. These high-angle mesoscale faults (Sharp *et al.* 2000) are antithetic to the major Loutraki and Heraion faults (Fig. 2), and possibly associated with the seismically active Saronic Gulf. Road cuttings along the steep access road to Agriliou Bay (Fig. 2) from the main Loutraki–Perachora road provide an excellent example of such displacement. Beach conglomerates are offset by 26 m on a 110°–290°N trending ‘minor’ fault. Uplift rates calculated from such a locally displaced palaeoshoreline will not be representative of the coastal section, and if the fault remained unidentified, the beach deposits on the footwall and hanging wall could be incorrectly interpreted as two separate highstand events. To be as rigorous as possible when determining spatial uplift rates we focus on shorelines that have been mapped over distances of hundreds of metres at similar elevations and are untilted by minor faults.

**Perachora peninsula: Holocene raised shorelines**

Marine notches cut into lithified coastal scarps are preserved on the west Perachora peninsula between Mylokopi and Cape Heraion (Fig. 2; Pirazzoli *et al.* 1994; Kershaw & Guo 2001; Leeder *et al.* 2003), and to just west of Agriliou Bay well-defined marine notches are mapped at 3.0–3.2 m. At many locations there is at least one additional notch at lower elevations, evidence of multiple uplift events. *Lithophaga* shells recovered from holes in the highest notch and dated by Pirazzoli *et al.* (1994) yield Holocene uplift rates of 0.50 ± 0.05 to 0.59 ± 0.1 mm a⁻¹.

At Agriliou Bay (Fig. 2) a reduction in uplift rates possibly marks a tectonic boundary. We interpret Leeder *et al.*’s (2007) 3.0 m shoreline at this location as a karstic surface rather than evidence of a marine platform. *Lithophaga* shells recovered from the highest notch at 1.6 m (*¹⁴C* dated to 6810–6727 cal. BP) mark a reduced uplift rate of 0.22 ± 0.01 mm a⁻¹. For 1 km east of Agriliou Bay prominent inner edges closest to sea level occur as notches at 1.5 and 0.3 m elevations, the latter with a 3–4 m wide abrasion platform (Fig. 5). Continuing east to Loutraki Bay (Fig. 2) we find that notches of similar preservation in comparable substrates are at 1.5 m or lower, until at Loutraki Bay the
highest notch is at 0.9 m. However, a *Lithophaga* shell recovered from 2.2 m near the top of a flight of bored holes yielded a 14C age of 2420 ± 40 cal. BP. The inner edge evident above the bored holes is defined by sea caves at 10 ± 0.5 m, which, if Holocene, is locally exceptional, as other Holocene palaeoshorelines on the Perachora peninsula are below 3.2 m elevation. The inner edge of MIS 5a age west of Loutraki Bay is at 10–12 m (see below; Fig. 4) consistent with other MIS 5a elevations in the area. Thus, we think it probable that the 14C age is unreliable and the 0.9 m notch is actually the highest Holocene inner edge at Loutraki Bay.

The majority of notches have a symmetrical profile, many with preserved delicate encrusting marine fauna that are considered reliable evidence of pulsed uplift events that rapidly lift the full depth of the notch above the reach of the normal tidal range so that it is protected from further erosion by the sea (Laborel & Laborel-Deguen 1994; Pirazzoli 2005).

**Perachora peninsula: pre-Holocene palaeoshorelines**

Table 1: U/Th analyses of *Cladochora caespitosa* stems sampled from in situ coral thickets

| Sample | Location | Elevation (m) | Age (ka) | Activity ratios | Uncertainties | Dissolution techniques |
|--------|----------|--------------|----------|----------------|--------------|------------------------|
| JS 34  | 5; 667579| 2.6          | 158 (3)  | 1.135 (5) 1.287 (10) | 5 ± 0.5 | Edwards et al. (1987) |
| JS 35  | 6; 668209| 2.6          | 158 (3)  | 1.135 (5) 1.287 (10) | 5 ± 0.5 | Edwards et al. (1987) |
| JS 36  | 7; 669257| 2.2          | 310 (4.9) | 1.287 (10) 1.551 (2) | 5 ± 0.5 | Edwards et al. (1987) |
| JS 37  | 8; 670001| 2.2          | 310 (4.9) | 1.287 (10) 1.551 (2) | 5 ± 0.5 | Edwards et al. (1987) |

Sample elevation and the correlated inner edge elevation are recorded. Activity ratios are shown with 1 SE uncertainty in parenthesis. Age uncertainties are 2 SE. Dissolution techniques have been described by Leeder et al. (1994) and Edwards et al. (1987). Decay constants for 234U, 238U, 230Th and 232Th are 2.82629E +06, 1.55125E +10, 9.15771E +06 and 4.9475E +11 a, respectively (Cheng et al. 2000).

Fig. 5. Raised marine notch (arrowed) cut in fissured Miocene limestone, with an inner edge at 1.5 m above modern high water and a fronting abrasion platform, 1 km east of Agriliou Bay. At this location the 0.3 m notch has been eroded. The photograph was taken at low tide; tidal range on this day was 0.75 m.

J. A. TURNER ET AL. 1242
preserved in situ corals at 10 m elevation, found in shoreface deposits, yield an MIS 5a age of 82.5 ± 2.1 ka (Table 1), supporting a previous date (70.2 ± 0.7 ka) by Dia et al. (1997). We map this MIS 5a shoreline along an inner edge at 12 m to the west side of the arroyo and on to an exposure 350 m to the west where beach conglomerates at 8.0–10 m elevation are preserved. These latter overlie corals dated to 81.9 ± 2.0 ka (Table 1), another MIS 5a age and in agreement with ages obtained by Roberts et al. (2009) at the same location. The average elevation for the inner edge is 10 ± 2 m, giving uplift rates since MIS 5a of 0.31 ± 0.42 mm a⁻¹.

A prominent terrace at 25–30 m at Makrugoaz Ridge was dated from coral samples by Leeder et al. (2003, 2005) and Roberts et al. (2009) to MIS 5e. This terrace is traceable to between Makrugoaz Ridge and Agriliou Bay at 25–30 m elevation, with occasional inner edge features at 32 m, a robust measure of uniform uplift of this coastal section at average rates of 0.18 ± 0.05 mm a⁻¹. A 27 m terrace inner edge at Agriliou Bay (Fig. 4) has proved much more difficult to date with confidence. Dia et al. (1997) obtained an MIS 5e age for corals at 8 m elevation in the sediment package below the 27 m inner edge. However, other studies on the same coal-bearing sediments (discussed below) have reported ages that range from MIS 5 to 9. It is important to critically review the age of this palaeoshoreline as Holocene uplift rates suggest a tectonic boundary at Agriliou Bay, expressed by change in uplift rate; we return to this later.

A MIS 7e shoreline in the adjacent ‘Australia Gorge’ (Figs 2 and 4) is reliably dated. Coral thickets are found in a fossiliferous unit traceable laterally to the back of a terrace where it is capped by beach conglomerates at 74 m inner edge. The chronology for this palaeoshoreline comes from three separate investigations. Corals present in cuttings on the west side of the road at 53 m are dated by us to 251.8 ± 19.4 ka (Table 1). These were previously dated by Dia et al. (1997) from 55 m elevation to 236.2 ± 3.0 and 250.1 ± 6.3 ka. Samples from the east side of the road were dated to 240 +5.7/-5.4 ka and 248 +7.2/-6.7 ka by Roberts et al. (2009). Using the elevation of the rigorously identified inner edge, the average uplift rate since MIS 7e is 0.31 ± 0.04 mm a⁻¹.

Between Heraion and Flagnoro Bay (Fig. 2) we also mapped terraces at consistent elevations of c. 45 m and c. 55 m; that is, between those dated to MIS 5e and MIS 7e (Fig. 4). We suggest that these terraces are contemporaneous and the c. 45 m shorelines are the same age as the terrace between Makrugoaz Ridge and Heraion dated by Leeder et al. (2005) to MIS 7a. The 55 m terrace is probably of MIS 7c age. Calculation of average uplift rates of the MIS 7a, 7c and 7e shorelines are 0.27 ± 0.04, 0.24 ± 0.04 and 0.31 ± 0.04 mm a⁻¹, respectively.

Above Agriliou Bay and Australia Gorge (Fig. 2), an undated marine terrace at 185 m is underlain by beach and shoreface sediments exposed in the west side of Panorama Gorge, the bedrock channel that feeds Agriliou Bay arroyo. The stratigraphy is offset by a flower structure (possibly a strike-slip pull-apart fault) described by Leeder et al. (2005, fig. 4). Preservation of this clearly defined terrace with its progradational stratigraphy suggests that it was formed over a prolonged highstand and demonstrates that uplift had been occurring for some time before MIS 7, possibly since c. 600 ka.

As uplift rates at Agriliou Bay are important (see above) we return to the issue of the age of the 27 m palaeoshoreline. The MIS 5e age (120.5 ± 0.9, 123.1 ± 0.7 ka) of coals at 8 m elevation (Dia et al. 1997) was contradicted by Roberts et al. (2009), who used careful pre-selective ‘preservation quality’ criteria including differential dating of coal septa and stem walls. Their results show that septa generally yielded older ages. Their samples from various units in the sediment package below the 27 m inner edge yielded ages between 161 ± 4.4 ka and 211 +4.2/-4.1 ka, which they assigned to MIS 7. A lower sample returned an age of 337 +37/-26 ka and was assigned to MIS 9. Using these dates Roberts et al. (2009) interpreted the sedimentology of the sediment package as an amalgam of MIS 9 to 5e age sediments with intervening lowstand surfaces. Our own attempts to date more than 30 coral stems from the same coal-bearing units using equally rigorous pre-screening criteria suggest that all our samples exhibit open-system geochemistry. We think it highly probable that the data published by Roberts et al. (2009) are affected from the same problem, explaining their range of ages, the older values being caused by diagenetic uranium loss. In the absence of reliable chronology we infer the age of the 27 m inner edge as MIS 5e by reference to the reliably dated 10 m MIS 5a shoreline and the 74 m elevation MIS 7 shoreline (discussed above). There is no evidence of faulting to complicate its elevation and an inferred MIS 5e age for the 27 m inner edge at Agriliou Bay suggests a mean uplift rate of 0.17 ± 0.04 mm a⁻¹.

East Agriliou Bay to Loutraki

From Agriliou Bay to Loutraki (Fig. 2) preserved palaeoshorelines are sparse but remnants cluster at 8–10 m, 22–26 m and c. 50 m elevation (Fig. 4). Coral samples exposed at 2 m in the eroding shoreline at Loutraki are dated to 137 ± 4.5 ka (Table 1). An associated inner edge has not been found, but the corals, and well-preserved marine shells in bioturbated sands to 4 m elevation, are good evidence that there has been no net subsidence of this coastal section since MIS 5e. The sea caves to which the dated Lithophaga shell (2420 ± 40 BP) is correlated, at 10 m elevation (see above), are eroded into fan breccia prograded over the caralliferous facies dated to MIS 5e. The sea caves were clearly cut by a highstand younger than c. 123 ka; MIS 5a or c is a likely candidate as shorelines at similar elevations are dated to MIS 5a further west (Fig. 4). If this chronology is correct, uplift rates are c. 0.3 mm a⁻¹. However, if the sea caves formed during a late MIS 5e sea-level highstand maximum of +6 to +9 m at c. 119 ka (Hearty et al. 2007), uplift rates are lower. Over longer time scales net average uplift has occurred, as undated marine platforms and neptunian dykes with beach sands and gravels are found at 43 and 70 m on the Lower Loutraki fault footwall above this exposure.

Loutraki plain

Loutraki plain sediments are alluvial sands and gravels with terra rossa soils that overlie raised brackish and marine sediments of the eastern Corinth basin. The north side of the plain is defined by the east–west-trending 800 m relief Lower Loutraki fault scarp (Figs 2 and 3); at the scarp base a bajada formed from coalesced alluvial fan, colluvium and scree deposits is incised by drainage from the south flank of the Gerania Range. The arroyo of most interest is that on the west side of the town cemetery, which reveals fan delta top and foresets off-lapping from c. 72 m over heavily bioturbated marine sandy marls. This palaeodelta has a terrace on its seaward margins inset at c. 53 m elevation with sea caves affording internal 3D views of the delta clinoforms. Although these elevations are similar to MIS 7 inner edges mapped to the west (c. 54 m and c. 74 m), correlation is speculative in the absence of chronology.
On the southern Loutraki plain, well-preserved and cemented beach gravels are exposed on the modern beach and are probably of Holocene age, but the inner edge is buried by spoil excavated from the Corinth canal. On the plain beside the Corinth to Loutraki road, and in shallow (<1.0 m) railway cuttings c. 3.5 km south of Loutraki town centre, intermittent and poor exposures of cemented marine sandy marls occur. Degraded oyster shells at c. 47 m elevation give clear evidence for uplift but there is no dateable fossil material or mappable inner edge with which to calculate uplift rates.

**Corinth plain and Isthmus**

The Isthmus (Fig. 2) is a land bridge that has existed during eustatic highstands since MIS 7 (Collier & Thompson 1991), and is now cut by the Corinth canal, the walls of which expose Pliocene and early Quaternary Corinthian marls capped by offlapping sequences of shallow marine gravels. Stretching west from the canal across the Corinth plain are flights of Late Quaternary marine terraces that parallel the coastline (Keraudren & Sorel 1987; Armijo et al. 1996; Fig. 6). Here, precise calculation of uplift rates is hindered by a lack of well-defined inner edge features. Palaeoshorelines below 4 m elevation are not well represented; typically they are obscured by the built environment or are poorly preserved. The best evidence from the south Lechaion Gulf coast comes from ancient Lechaion port (Fig. 2). Here, the raised walls of the ancient port are *Lithophaga* bored at 1.1 m; archaeological evidence also suggests an initial uplift event followed by subsidence, inferred from submergence of the outer harbour moles to < −0.7 m (Flemming et al. 1973), an interpretation supported by Vita-Finzi (1993). We mapped cemented beach gravels to 2.0 m beside the main coastal road inland of the ancient Lechaion archaeology site but found no datable material.

The marine sequences of the New Corinth Terrace (Fig. 6), clearly exposed west of modern Corinth beside the coastal road, were dated to MIS 5ε (Leeder et al. 2005). This terrace is backtilted to the south (c. 5 m footwall uplift) with no inner edge features, but the terrace back is at c. 30 m elevation, giving uplift rates of c. 0.22 ± 0.05 mm a⁻¹. North of Examilia (Fig. 2), the ‘Old Corinth’ terrace (Fig. 6) at c. 75 m was dated to MIS 7 by Armijo et al. (1996) using Collier & Thompson’s (1991) coral ages of 232.0 +23.6/-19.7 ka and 177.7 +11.0/-10.1 ka. This chronology was contradicted by Dia et al. (1997), who sampled coral stems from the same outcrop, 1.5 km north of Examilia, and obtained MIS 9 ages of 329.7 ± 7.4 ka to 385.5 ± 21.4 ka, a problem we return to below.

Near Examilia, quarried oolitic sand dunes lie on the hanging wall to the Examilia minor fault and a second fault to the SE that Collier & Thompson (1991) proposed was a submarine scarp essential for acceleration of tidal currents to build the oolitic dunes during MIS 7 (Figs 2 and 6). This chronology was determined by assuming that the dunes were contemporaneous with their MIS 7 corals. As the fault footwall (now a terrace at...
The activity status of basin bounding faults

With uplift rates determined for locations around the Lechaion Gulf margins it remains to examine possible mechanisms for this uplift, particularly the activity status of basin bounding faults and their contribution to uplift or subsidence of mapped features. The palaeoshorelines where uplift rates are critical for model testing are on the footwall and hanging wall of crustal-scale faults bounding the Lechaion Gulf; these are the south-dipping Lower Loutraki, Upper Loutraki, Atlae and Heraion faults and the north-dipping Kenchriea fault (Fig. 2). We also discuss a proposed west extension of the Pisia fault from Perachora to Lake Vouliagmeni (dashed line, Fig. 2), suggested as the mechanism for raising the Perachora south coast between Lake Vouliagmeni and Agriiou Bay (Morewood & Roberts 1999). An important observation is that the Lechaion Gulf is deepest on its northern side, suggesting greater activity of the south-dipping faults; these have a greater slip rate and/or are active over a longer period than the Kenchriea fault that bounds the southern basin margin, but the current activity of all faults is uncertain.

Goldsworthy & Jackson (2001) proposed an active Lower Loutraki fault, which they projected westward, continuing offshore on the fault segments we define as the Atlae and Heraion faults (Fig. 2); slip on these faults must contribute to uplift of the Perachora peninsula. Leeder et al. (2005) concluded that the Lower Loutraki fault is probably Holocene inactive, and noted that the Upper Loutraki fault requires further study. Goldsworthy & Jackson (2001) also considered the south basin bounding Kenchriea fault to be active, in agreement with others (e.g. Roberts & Jackson 1991; Armijo et al. 1996; Moretti et al. 2003) but we re-examine the evidence for Holocene displacement of this fault, as the degraded fault scarp and wide bajada appear inconsistent with an active status.

Lower Loutraki fault

The onshore Lower Loutraki fault (Fig. 2) trends east–west, defined by the degraded south face of the Gerania Range. The base-of-scarp bajada is gullied by channels that drain the c. 1000 m high range. We find no evidence for major displacement of Holocene fan breccia and colluvium. The marine sediments in its hanging wall on the southern Loutraki plain (at 47 m) and bajada (72 m) discussed above are clearly uplifted, which suggests that the fault has not been recently active, possibly not since at least MIS 7. However, an end of activity of the fault is not conclusive as kilometre-scale faults may rupture on time scales that exceed preservation of displacement in the surface expression of drainage and sediments. The Kaparelli fault, for example, in the NW Alkyonides Gulf was inactive for 10 ka before the 1981 rupture (Benedetti et al. 2003).

Upper Loutraki and Atlae faults

There is evidence for Holocene activity on the 3 km long structure we name the Upper Loutraki fault (Fig. 2). The Holocene colluvium wedge in its immediate hanging wall exposed by quarrying is dragged upward adjacent to the fault scarp, deformed by footwall slip. The upper 1.0–4.0 m of a very marked fault scarp in basement limestone is clearly visible across the front of the range at c. 300 m elevation, and although degraded the escarpment above is markedly more so. Subsidence of the Upper Loutraki hanging wall is accommodated at the east tip by a north–south fault-controlled ravine, the east coast of the Lechaion Gulf coincident with this trend, marking the transition

120 m) had to be submerged, the corresponding inner edge for the MIS 7 highstand must be at least at this modern elevation (120 m). An alternative terrace chronology proposed by Armijo et al. (1996) correlated the MIS 7 coral age only with the c. 75 m elevation Old Corinth terrace (Fig. 6), and assigned an MIS 9 age to the oolitic dunes (‘Temple’ terrace) which, using Collier & Thompson’s (1991) dune-building process, has a 120 m inner edge. The chronology of Armijo et al. (1996) involves a sequence of terraces increasing in age with higher elevation. Collier & Thompson (1991) required three terraces (75 m, 85 m and 120 m) to be of MIS 7 age. The terrace ages after Dia et al. (1997) allocated both the Old Corinth and Temple terraces to MIS 9, with no MIS 7 terrace. It is reasonable to suspect that the coral ages of the 75 m Old Corinth terrace determined by Dia et al. (1997) were older because of open-system conditions, and that the MIS 7 age obtained by Collier & Thompson (1991) is plausible; this gives uplift rates for the MIS 7 Old Corinth of 0.17 ± 0.05 mm a⁻¹. The Temple terrace and oolitic dunes with the inner edge with the 120 m elevation are probably of MIS 9 age yielding average uplift rates of 0.39 +0.1/−0.09 mm a⁻¹. We add an uncertainty of +0/−0.09 mm a⁻¹ to this age in case of post-depositional footwall uplift (≤20 m since 320 ka) on the Examilia fault and an uncertain inner edge elevation (+10 m).

The east margin of the Corinth plain is bounded by the tectonically active Saronic Gulf (Fig. 2). Between the Corinth canal and Kechrion village to the south, a series of low-lying deltas and interfluvies of semi-lithified red sands and gravels crop out. Sparse exposures of marine sediment occur 0.5 km south of the Corinth canal at similar elevations to the MIS 7e corals recorded in the canal walls by Collier (1990). At the southern limit of the study area, the Ancient Kenchriea port near modern Kechrion village has clearly subsided and the port buildings are now below sea level; historically documented subsidence associated with earthquake events was reported by Scraton et al. (1978). The lateral extent of subsidence is uncertain but may be limited to the harbour area as archaeoological studies of Roman burial chambers c. 0.5 km north of the sunken harbour are not disturbed by tectonic activity (Sarris et al. 2006).

Within the Corinth canal (Figs 2 and 6), marine sequences studied in detail by Neuser (1982), Collier (1990) and Richter & Neuser (1998) were dated to MIS 7, 9 and >9 by Collier (1990). Collier (1990) discussed the interplay of syntectonic deposition of a variable sediment supply and suggested that the depositional geometries favour eustatic sea-level rise exceeding that of tectonic uplift. The highest elevation marine sequence in the canal is at c. 80 m on the central horst, interpreted as a MIS 11 highstand by Collier (1990). This is onlapped by younger highstand deposits with an MIS 9 sequence to 70 m and MIS 7 to 47 m (Collier et al. 1992, fig. 2a) that yield average uplift rates of 0.22 ± 0.04 and 0.19 ± 0.04 mm a⁻¹, respectively. These uplift rates are much lower than those near Examilia, implying an increase in net uplift west of the Isthmus, in agreement with Armijo et al. (1996). The canal walls clearly show offset by minor faults that have no surface expression, and faults dip seaward on either side of the central horst, with offsets from <1 to c. 40 m. We check the contribution to uplift of the c. 80 m elevation central horst by these faults. The calculations assume that there is no decay in uplift on the footwall away from the fault line, and all faults stepped sequentially away from the horst. This result is an absolute maximum net uplift of 60 m, and true uplift is expected to be less. The minor faults cannot explain uplift of the Isthmus to 80 m elevation, so another tectonic uplift mechanism must be active here.
from the deep-water Loutraki port to the town’s promenade and beaches.

The offshore fault we refer to as the Atlae segment of the Lower Loutraki system parallels the Upper Loutraki fault (Fig. 2) and is of uncertain activity status. The raising of palaeoshorelines requires an additional 100 m of unexplained uplift (section b–b’, Fig. 3a); this may be by an active Atlae fault, and nearshore seismic survey is needed to determine its activity status.

**Heraion and Perachora offshore faults**

These are defined by steep coast-parallel bathymetric gradients offshore of the western Perachora peninsula (Fig. 2) and define the geometrical form of the peninsula. The Heraion fault is active because the results from offshore seismic surveys appear to show cliniforms onlapping the footwall scarp that is dragged upwards (Perissoratis et al. 1986; Sakellariou et al. 2004; D. Sakellariou, pers. comm.; J. Weiss, pers. comm.). The profiles of the notches on the Perachora peninsula are typically interpreted as pulsed uplift (e.g. Laborel & Laborel-Deguen 1994; Pirazzoli et al. 1994; Leeder et al. 2005), which we mapped to at least 3 m at frequent intervals between Heraion and Agriliou Bay, but decreasing in elevation at Agriiliou Bay. This onshore evidence points to incremental co-seismic uplift by an active offshore fault that tips out just east of Agriliou Bay.

The Perachora fault (Fig. 2) is determined as possibly active from a seismic survey by Stefatos et al. (2002, fig. 7; A. Stefatos, pers. comm.) but the seismic profile of dragged hanging-wall reflectors is rather indistinct. However, I. Moretti (pers. comm.) interpreted an active status of the Perachora fault from a seismic line of Moretti et al. (2004, fig. 2). This is supported by marine notches to c. 3 m on the footwall to the fault (Pirazzoli et al. 1994) that are good evidence of co-seismic footwall uplift.

**Kenchriae fault**

Noller et al. (1997) described a c. 1.5 m height scarplet that cuts Holocene sediments onlapping the deeply gullied and degraded Kenchriae fault escarpment (Fig. 2), as evidence of a Holocene fault rupture. Goldsworthy & Jackson (2001) concurred, albeit at low slip rates, and attributed subsidence of the ancient Kenchriae port at Kechries (Fig. 2) to co-seismic hanging-wall subsidence. However, the scarplet is limited in extent to clearance of natural vegetation and is coincident with an orchard on the slopes below. We interpret this feature as an agricultural terrace and from this, together with the highly degraded state of the fault scarp and well-devolved bajada, infer that the Kenchriae fault is inactive.

Subsidence of Kenchriae harbour was probably on the footwall of a local minor fault just to the NE (Fig. 2), previously identified by Collier & Thompson (1991).

**Western Pisia fault strand**

This onshore north-dipping fault strand between Makrugoaz Ridge and Perachora village (Fig. 2) has been proposed to explain field interpretations of uplift rates of 0.29 mm a·¹ at Makrugoaz Ridge increasing to 0.55 mm a·¹ at Agriiliou Bay (Morewood & Roberts 1999; Cooper et al. 2007; Roberts et al. 2009). However, we contend that there has been no recent activity of this fault. The north-facing limestone and serpentinite outcrops that define the fault scarp are much degraded along its length; eroded material forms fans that grade from midway up the fault line, contrasting with the steeply dipping active Pisia and Skinos faults just to the east. The fissures in the vegetated bajada identified by Roberts et al. (2009) as evidence of slip in 1981 are here interpreted to be the result of slope failure, possibly caused by ground shaking during earthquake events such as the 1981 rupture of the Pisia fault (Jackson et al. 1982).

The elevation of the MIS 5e shoreline was mapped to 70 m (which we also dispute; see the reference above to the 27 m inner edge) by Roberts et al. (2009, fig. 6) on the footwall of this fault. Such uplift is expected to require a hanging-wall subsidence of at least 140 m (footwall:hanging wall partition of 1:2) giving a total displacement since c. 123 ka of 210 m; the calculated offset (from 1:5000 topographic maps) is just 100 m. We thus find no compelling evidence of late Pleistocene or Holocene activity along this western Pisia fault strand and do not consider this fault as contributing to uplift of the Perachora south coast or subsidence of the north coast. However, we welcome defining evidence, perhaps obtainable by trenching across the fault line (see Collier et al. 1998).

**Discussion of uplift models**

The footwall uplift hypotheses outlined above should result in spatial gradients in the elevation of same-age shorelines uplifted around the Lechaion Gulf, whereas the isostatic ‘regional’ uplift model predicts uniform rates (Fig. 3). To determine the contribution of fault-controlled (strain-related) and isostatic uplift that explain the present elevation of the Lechaion Gulf margins we compare observed uplift rates with those predicted, and evaluate evidence for the mode of uplift. In Figure 6 we add the observed marine terraces and sampling points of dated corals and Lithophaga to the map. The cross-sections show the observed inner edge elevations, and on the graph below each cross-section we transfer the model predictions from Figure 3 and represent the observed uplift rates with continuous-line segments.

To quantify observed uplift rates we use data only from shorelines mapped at the same elevations over distances of hundreds of metres that are unlikely to have been disturbed by minor faults. The elevations of observed Holocene, MIS 5e and MIS 7 shorelines around the Lechaion Gulf are summarized in Table 2, and compared with the model predictions as summarized in Figure 6.

The Holocene notches of the Perachora peninsula (Figs 4 and 5) clearly indicate pulsed uplift of the west peninsula at rates of at least 0.50 ± 0.05 mm a·¹, slowing to 0.22 ± 0.01 mm a·¹ at Agriiliou Bay, which probably marks the east tip of the active offshore Heraion fault. Shoreline displacement rates between Agriiliou Bay and Loutraki are not constrained accurately but pulsed uplift is evident from the raised notch profiles. A lack of exposed Holocene inner edges on the Corinth and Loutraki plains does not allow quantification of uplift rates, but historical documents record pulsed uplift and subsidence near Lechaion; we find no means to differentiate between movement by minor faults that bisect this area and that attributable to the long-term uplift.

The well-preserved MIS 5e (c. 123 ka) shoreline on the northern coast of the Lechaion Gulf is a ubiquitous uplift marker for the western Perachora peninsula (Fig. 6). Uplift rate of MIS 5e shorelines (0.18 ± 0.05 mm a·¹) is slower than the rates of Holocene and MIS 7 shorelines (Table 2). As the western tip of the Perachora peninsula is not on the hanging wall of any faults, but undergoes uplift only on the footwalls of the Perachora and Heraion faults (Figs 2 and 6) the reduced average uplift rate of MIS 5e shorelines indicates a change in slip rates of these faults between the Holocene and MIS 7. The notches at 32 m on the Perachora peninsula (Fig. 4) demonstrate that co-seismic uplift.

...
also occurred during the last interglacial at very similar elevations of 27–32 m. The MIS 5e uplift rate is slightly faster (0.25 ± 0.08 mm a⁻¹) on the southern Lechaion Gulf, as shown by the New Corinth Terrace elevation (Figs 2 and 6).

On the Perachora peninsula the MIS 7e shoreline elevation and age gives a secure uplift rate of 0.31 ± 0.04 mm a⁻¹ at ‘Australia Gorge’ (Figs 2 and 6; Table 2). On the southern margin at some distance from the active faults this reduces to 0.19 ± 0.04 mm a⁻¹ at the Corinth canal but terrace elevation increases west along the Corinth plain (Armijo et al. 1996). There are no dated MIS 9 shorelines on the Perachora peninsula. If the Temple terrace of Armijo et al. (1996) is MIS 9 age it has been uplifted at 0.39 +0.01/-0.09 mm a⁻¹, faster that the MIS 9 shoreline at the Corinth canal, which was raised at 0.19 ± 0.04 mm a⁻¹, again supporting a westward increase in average uplift rates.

The cross-sections in Figure 6 show that observed uplift (continuous-line segments) for the west Perachora peninsula (section a–a’) is less than that predicted by fault slip; either there is a small component of isostatic uplift or the assumptions made to add quantitative data to the models resulted in slightly higher predicted values. The agreement with the ‘regional-scale’ model (Fig. 6, section c–c’) is not surprising, as Leeder predicted values. The agreement with the ‘regional-scale’ model resulted in slightly higher isostatic uplift or the assumptions made to add quantitative data to the models resulted in slightly higher isostatic uplift contributions. The isostatic uplift model of Leeder et al. (2003) better fits the observed data. However, those workers predicted uniform uplift across the region; instead, the elevation of MIS 7 and 9 shorelines increases westwards across the Corinth plain from the Corinth canal on the Isthmus to Examilia. We find no conclusive evidence to determine whether the isostatic uplift is pulsed or steady.

As normal faults are active on the Perachora peninsula we need to determine whether the Lechaion Gulf is extending and thus still part of the active rift system. Avallone et al.’s (2004, fig. 3) geodetic measurements of Aegean plate motion included a GPS network datum on the Perachora peninsula. This allows the geodetic motion across the Lechaion and Alkyonides Gulfs to be separated. Scaled velocity vectors across the western Lechaion Gulf show the southern part moving more rapidly to the SW, resulting in extension across the western Lechaion Gulf of 4 mm a⁻¹, and across the eastern end at 2 mm a⁻¹. These rates are high; if we assume that the Heraion fault takes up strain between Cape Heraion and the Corinth plain, and dips at 45° so that extension equals vertical displacement, then maximum extension can be only 0.9 mm a⁻¹ (0.3 mm a⁻¹ uplift with a footwall-hanging wall partition of 1:2). There are two smaller offshore faults at the mouth to the Lechaion Gulf (Fig. 2) (Sakellariou et al. 2004, fig. 2) where a component of the missing strain may be taken up. However, the 2 mm a⁻¹ extension in the east Lechaion Gulf is not explained. Further measurements of strain across the Lechaion Gulf are required to resolve these problems. It is, however, likely that the fault system of the Perachora peninsula (Psia, Skinos, Perachora and Heraion faults) is taking up strain in a transfer zone between the East Alkyonides faults to the SE and the Xylocastro fault of the Gulf of Corinth to the SE (Figs 1 and 2) and thus making a small contribution to rifting.

To set isostatic uplift in the context of the wider rift system we compare uplift of the Lechaion Gulf sediments with that of the Megara Basin proto-rift basin, some 28 km to the east, and the Gulf of Corinth to the west (Fig. 2). Uplift of the Megara Basin began in the late Pliocene about 2.2 Ma (Leeder et al. 2008) and continued into the late Pleistocene (Bentham et al. 1991). Here, uplift declines with distance from the footwall of the East Alkyonides and Psatha faults to zero at the Saronic Gulf coast (Collier et al. 1992), suggesting that any isostatic uplift at this site is a small contribution to rifting. West of the Lechaion Gulf, Rohais et al. (2008) reported an unexplained
component of uplift in the centre of the Gulf of Corinth southern margin and McNeill et al. (2005) suggested that Leeder et al.’s (2003) ‘regional uplift’ has expression in the Elikia area.

The data presented here broadly support uplift of the Perachora peninsula by fault slip (Jackson et al. 1982; Armijo et al. 1996) and the notion that isostatic effects cause uplift of the Isthmus and Corinth plain at least (Leeder et al. 2003, fig. 4). Leeder et al. suggested that the mechanism for uplift is caused by isostatic adjustment of the lithosphere as over-thickened crust pushes Aegea over the shallow subducted African slab under the Gulf of Corinth. This requires the position of the African hinge from shallow to steep subduction to be under the Gulf of Corinth (Tibiri et al. 2000) rather than further south under the Argolikos Gulf (Papazachos et al. 2000; Zelt et al. 2005). The position of the Gulf of Corinth relative to the Hellenic subduction zone and the subducting African slab (Fig. 1) suggests that lithospheric and upper mantle processes in the study region may be complex. Uplift by positive thermal buoyancy from a warmed lithosphere is discounted in the absence of a thermal anomaly under the Gulf of Corinth (Moretti et al. 2003). Faccenna & Becker (2010) modelled vigorous small-scale convection in mobile belts and used the eastern Mediterranean as an example to show how this may result in a coupling between the deeper lithosphere and upper mantle. Their model shows that push of Aegea to the SW is probably dominated by crustal gravitational potential energy of the thickened Aegean plate, broadly supporting the lithospheric dynamics of Leeder et al. (2003).

Conclusion

Spatial uplift rates calculated for raised shorelines from the north and south Lechaion Gulf margins have been used to test various models (Jackson et al. 1982; Armijo et al. 1996; Leeder et al. 2003) that attempt to explain uplift dynamics. The northern margin of the Lechaion Gulf, the Perachora peninsula, is being uplifted on the footwall of the Pisia, Skinos, Upper Loutraki and offshore Heraion and Perachora faults at average rates of 0.31 ± 0.04 mm a⁻¹ since MIS 7. We also find that uplift is variable over orbital time scales.

The southern margin of the Lechaion Gulf is being exhumed independent of fault slip, and uplift of basin sediments is attributed to the isostatic uplift model of Leeder et al. (2003). However, uplift is not spatially uniform. The westward gradient in uplift is best shown by the MIS 7 shoreline, for which uplift rates increase from 0.19 ± 0.05 mm a⁻¹ at the Corinth canal to 0.31 ± 0.05 mm a⁻¹ at Examilia.

This study identifies, and distinguishes between, uplift by fault slip and uplift by isostatic adjustment in a young continental rift setting. We have clearly demonstrated and quantified a component of uplift that is explained by lithospheric adjustment. These findings are relevant to calculations of fault slip, and thus extension rates, in the Gulf of Corinth rift. The contribution of isostatic uplift to footwall elevation should be recognized to prevent overestimation of uplift rates by faults and thus the magnitude of fault slip. The recognition of isostatic uplift should also contribute to better understanding and modelling of the overriding lithosphere in equivalent subduction-zone settings.

We thank NERC for funding U-Th dating of corals by the OUUSF (IP/961/0507), and the School of Environmental Sciences, University of East Anglia, for funding J.A.T.’s PhD. G. Mack kindly provided funds for 14C dating Lithophaga shells. The Institute of Geology and Mineral Exploration (IGME) gave us permission to continue our research in Greece. For their helpful and stimulating discussions we thank G. Mack and V. Bense.

D. Sakellariou and J. Weiss shared their latest ideas on the activity of offshore faults with us. An earlier version of the paper was greatly improved by the helpful comments of J. Dixon and I. Moretti.

References

Armeo, R., Meyer, B., King, G.C.P., Rigo, A. & Papannastassiou, D. 1996. Quaternary evolution of the Corinth Rift and its implications for the Late Cenozoic evolution of the Aegean. Geophysical Journal International, 126, 11–53.

Avalone, A., Broile, P., et al. 2004. Analysis of eleven years of deformation measured by GPS in the Corinth Rift Laboratory area. Comptes Rendus Geosciences, 336, 301–311.

Bell, R.E., McNeill, L.C., Bull, J.M., Hensstock, T.J., Collier, R.E.L. & Leeder, M.R. 2009. Fault architecture, basin structure and evolution of the Gulf of Corinth rift, central Greece. Basin Research, 21, 824–855.

Benedetti, L., Finkel, R., et al. 2003. Motion on the Kaparelli fault (Greece) prior to the 1981 earthquake sequence determined from 36Cl cosmogenic dating. Terra Nova, 15, 118–124.

Bentham, P., Collier, R.E.L., Gawthorpe, R.L., Leeder, M.R., Prossor, S. & Stark, C. 1991. Tectonosedimentary development of an extensional basin—the Neogene Megara Basin, Greece. Journal of the Geological Society, London, 148, 923–934.

Billiris, H., Paradissis, D., et al. 1991. Geodetic determination of tectonic deformation in central Greece from 1900 to 1988. Nature, 350, 124–129.

Broile, P., Rigo, A., et al. 2000. Active deformation of the Corinth rift, Greece: Results from repeated global positioning system surveys between 1990 and 1995. Journal of Geophysical Research B: Solid Earth, 105, 25605–25625.

Cheng, H., Edwards, R.L., Hoff, J., Gallup, C.D., Richards, D.A. & Asmerom, Y. 2000. The half-lives of uranium-234 and thorium-230. Chemical Geology, 169, 17–33.

Clarke, P.J., Davies, R.R., et al. 1997. Geodetic estimate of seismic hazard in the Gulf of Korinthos. Geophysical Research Letters, 24, 1303–1306.

Collier, R.E.L. 1990. Eustatic and tectonic controls upon Quaternary coastal sedimentation in the Corinth Basin, Greece. Journal of the Geological Society, London, 147, 301–314.

Collier, R.E.L. & Thompson, J. 1991. Transverse and linear dunes in an Upper Pleistocene marine sequence, Corinth Basin, Greece. Sedimentology, 38, 1021–1040.

Collier, R.E.L., Leeder, M.R., Rowe, P.J. & Atkinson, T.C. 1992. Rates of tectonic uplift in the Corinth and Megara Basins, Central Greece. Tectonics, 11, 1159–1167.

Collier, R.E.L., Pantorti, D., Addeggio, G., Di Martini, P.M., Masema, E. & Sakellaridis, D. 1998. Paleosesimetricity of the 1981 Corinth earthquake fault: Seismic contribution to extensional strain in central Greece and implications for seismic hazard. Journal of Geophysical Research, 103, 30001–30019.

Cooper, F.J., Roberts, G.P. & Underwood, C.J. 2007. A comparison of 10⁶–10⁶ year uplift rates on the South Alkomydes Fault, Central Greece: Holocene climate stability and the formation of coastal notches. Geophysical Research Letters, 34, L14310, doi:10.1029/2007GL030673, 032007.

Daradich, A., Mitrovica, J.X., Pysklywec, R.N., Willett, S.D. & Forte, A.M. 2003. Mantle flow, dynamic topography, and rift-flank uplift of Arabia. Geology, 31, 901–904.

Dia, A.N., Cohen, A.S., O’Nions, R.K. & Jackson, J.A. 1997. Rates of uplift investigated through 230Th dating in the Gulf of Corinth (Greece). Chemical Geology, 138, 171–184.

Ebbing, C.J., Jackson, J.A., Foster, A.N. & Hayward, N.J. 1999. Extensional basin geometry and the elastic lithosphere. Philosophical Transactions of the Royal Society, Series A, 357, 741–765.

Edwards, R.L., Chen, J.H., Ku, T.L. & Wassenburg, G.J. 1987. Precise timing of the last interglacial period from mass spectrometric determination of thorium-230 in corals. Science, 236, 1547–1553.

Faccenna, C. & Becker, T.W. 2010. Shaping mobile belts by small-scale convection. Nature, 465, 602–605.

Flemming, N.C., Czartoryski, N.M.G. & Huneter, A.M. 1973. Archaeological evidence for eustatic and tectonic components of relative sea level change in the south Aegean. Colston Symposium Papers, 23.

Goldsworthy, M. & Jackson, J. 2001. Migration of activity within normal fault systems: examples from the Quaternary of mainland Greece. Journal of Structural Geology, 23, 489–506.

Gray, J.M. & Ivanovich, M. 1988. Age of the Main Rock Platform, Western Scotland. Palaeogeography, Palaeoclimatology, Palaeoecology, 68, 337–345.

Hatzfeld, D., Martinod, J., Basteit, G. & Gautier, P. 1997. An analog experiment for the Aegean to describe the contribution of gravitational potential energy. Journal of Geophysical Research B: Solid Earth, 102, 649–659.

Hearty, P.J., Hollin, J.T., Neumann, A.C., O’Leary, M.J. & McCulloch, M.
The Evolving Continents: Understanding the Processes of Continental Growth
Edited by T. M. Kusky, M.-G. Zhai and W. Xiao
This Special Publication of the Geological Society of London, The Evolving Continents: Understanding Processes of Continental Growth, was written to honour the career of Brian F. Windley, who has been hugely influential in helping to achieve our current understanding of the evolution of the continental crust, and who has inspired many students and scientists to pursue studies on the evolution of the continents. Brian has studied processes of continental formation and evolution on most continents and of all ages, and has educated and inspired two generations of geologists to undertake careers in studies of continental evolution. The volume is organized into six sections, including: oceanic and island arc systems and continental growth; tectonics of accretional orogens and continental growth; growth and stabilization of continental crust: collisions and subduction processes; Precambrian tectonics and the birth of continents; and active tectonics and geomorphology of continental collision and growth zones.

Continental Tectonics and Mountain Building: The Legacy of Peach and Horne
Edited by R. D. Law, R. W. H. Butler, R. E. Holdsworth, M. Krabbendam and R. A. Strachan
The world's mountain ranges are the clearest manifestations of long-term deformation of the continental crust. As such they have attracted geological investigations for centuries. Throughout this long history of research a few keynotes publications stand out. One of the most important is the Geological Survey's 1907 Memoir on The Geological Structure of the North-West Highlands of Scotland. The Memoir summarized some of the Geological Survey's finest work, and outlined many of the principles of field-based structural and tectonic analysis that have subsequently guided generations of geologists working in other mountain belts, both ancient and modern. The thematic set of 32 papers in this Special Publication celebrate the 100th anniversary of the 1907 Memoir by placing the original findings in both historical and modern contexts, and juxtaposing them against present-day studies of deformation processes operating not only in the NW Highlands, but also in other mountain belts.

Advances in Interpretation of Geological Processes: Refinement of Multi-scale Data and Integration in Numerical Modelling
Edited by M. I. Spalla, A. M. Marotta and G. Gosso
Iterative comparison of analytical results and natural observations with predictions of numerical models improves interpretation of geological processes. Further refinements derive from wide-angle comparison of results from various scales of study. In this volume, advances from field, laboratory and modelling approaches to tectonic evolution - from the lithosphere to the rock scale - are compared. Constructive use is made of apparently discrepant or non-consistent results from analytical or methodological approaches in processing field or laboratory data, P-T estimates, absolute or relative age determinations of tectonic events, tectonic unit size in crustal-scale deformation, grain-scale deformation processes, various modelling approaches, and numerical techniques.

Tectonic and Stratigraphic Evolution of Zagros and Makran during the Mesozoic-Cenozoic
Edited by P. Leturmy and C. Robin
The Zagros fold-thrust belt (ZFTB) extends from Turkey to the Hormuz Strait, resulting from the collision of the Arabian and Eurasian plates during Cenozoic times, and separates the Arabian platform from the large plateaux of central Iran. To the east a pronounced syntaxis marks the transition between the Zagros collision belt and the Makran accretionary wedge. In the ZFTB, the Proterozoic to Recent stratigraphic succession pile is involved in huge folds, and offers the opportunity to study the stratigraphic and tectonic evolution of the Palaeo-Tethyan margin.

Few recent data were widely available on the southern Tethys margin preserved in the Zagros Mountains. The Middle East Basins Evolution (MEBE) program was an excellent opportunity to go back to the field and to collect new data to better constrain the evolution of this margin. In this volume the structure of the Zagros Mountains is explored through different scales of study. In this volume, advances from field, laboratory and numerical approaches are compared. Constructive use is made of apparently discrepant or non-consistent results from analytical or methodological approaches in processing field or laboratory data, P-T estimates, absolute or relative age determinations of tectonic events, tectonic unit size in crustal-scale deformation, grain-scale deformation processes, various modelling approaches, and numerical techniques.