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Long-term versus short-term deformation of the meizoseismal area of the 2008 Achaia–Elia (M_W 6.4) earthquake in NW Peloponnese, Greece: Evidence from historical triangulation and morphotectonic data

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ABSTRACT

The deformation of the meizoseismal area of the 2008 Achaia–Elia (M_W 6.4) earthquake in NW Peloponnese, of the first significant strike slip earthquake in continental Greece, was examined in two time scales; of 10^2 years, based on the analysis of high-accuracy historical triangulation data describing shear, and of 10^5–10^6 years, based on the analysis of the hydrographic network of the area for signs of streams offset by faulting. Our study revealed pre-seismic accumulation of shear strain of the order of 0.2 μrad/year in the study area, consistent with recent GPS evidence, but no signs of significant strike slip-induced offsets in the hydrographic network. These results confirm the hypothesis that the 2008 fault, which did not reached the surface and was not associated with significant seismic ground deformation, probably because of a surface flysch layer filtering high-strain events, was associated with an immature or a dormant, recently activated fault. This fault, about 150 km long and discordant to the morphotectonic trends of the area, seems first, to contain segments which have progressively reactivated in a specific direction in the last 20 years, reminiscent of the North Anatolian Fault, and second, to limit an 150 km wide (recent?) shear zone in the internal part of the arc, in a region mostly dominated by thrust faulting and strong destructive earthquakes.

1. Introduction

The Aegean and the Continental Greece are notable for their active normal faulting and the associated earthquakes (Mercier et al., 1979; Shaw and Jackson, 2010) and only a zone of thrusting is observed to the NW, along the Ionian Sea front, while evidence of strike slip faulting inlands in the mainland and the Aegean Islands is at least meager (Mercier et al., 1979; Underhill, 1989). For this reason, the 2008 M_W 6.4 seismic sequence in Achaia–Elia near Patras, in SW Greece (Fig. 1) was a surprise, because it was the first significant earthquake providing clear seismological evidence of strike slip faulting in this region. This was not the only surprise. This earthquake was not associated with any known fault; on the contrary it was discordant with the neotectonic trends of the area. Furthermore, in contrast to the predictions of elastic modeling, it produced no significant ground deformation, especially afterslip, as GPS and INSAR data reveal (Briole et al., 2008; Feng et al., 2010; Ganas et al., 2009; Giannopoulos et al., 2013; Margaris et al., 2010; cf. Marone et al., 1991).

These led Feng et al. (2010) to assign the 2008 earthquake to an immature fault, and the absence of surface deformation to thin-skin tectonics (cf. Fialko et al., 2005), imposed by a decollement horizon associated with low-strength flysch deposits (cf. Bernard et al., 2006; Kamberis et al., 2000) not permitting the propagation of the rupture to the surface.

While the available data permit to understand the deformation of the area during the 2008 earthquake, no information on the long-term and the pre-seismic deformation is available, especially on the process of accumulation and release of shear strain.

In order to shed some light on this problem, the tectonic deformation of the area in two time scales was examined. In a scale of ~100 years, based on the analysis of historical triangulation data, and in a scale of 10^5–10^6 years, based on the analysis of the hydrographic network of the area, focusing on possible stream offsets, indicative of unknown major historical or prehistoric strike slip earthquakes.

This study is important (1) to confirm or revise the hypothesis of an immature 2008 fault, (2) to put some constraints in the process of shear strain build up along the trace of the 2008 fault, (3) to discuss the relationship between shear and thrust deformation along the

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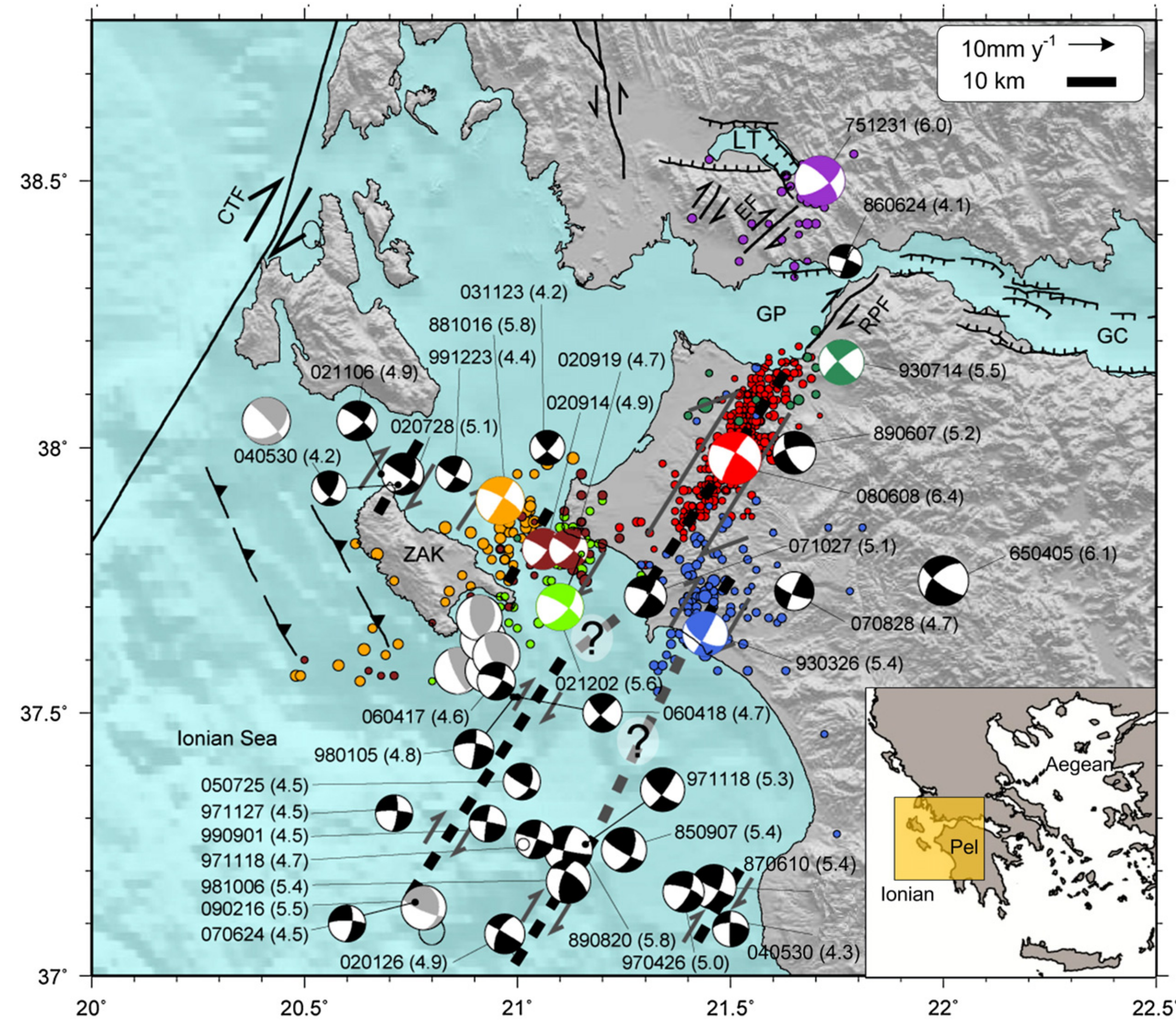

Fig. 1. Seismicity and focal mechanisms in the wider region of the 2008 earthquake in NW Peloponnese, Greece (red symbols; other colors indicate different sources of data, see Feng et al., 2010 for details). The strike slip fault offshore Cephalonia (CTF) and the possible strike slip fault marking the east edge of the Patras Gulf (PRF) are shown. This last fault seems to delineate with the predicted trace of the 2008 fault and those of previous strike slip earthquakes farther SW, indicating an about 150 km long immature or dormant fault, segments of which reactivated in a NE direction in the last 20 years. A zone of shear between these faults is inferred. Location map in the inset.
Modified from Feng et al. (2010).

Aegean Arc, and (4) to shed some light on the problem of seismic risk of Patras, the major city (approximately 250,000 inhabitants) in western Greece, mainly because of the apparent, gradual propagation of seismic faulting towards Patras (Fig. 1).

2. Geological and geodynamic background

Various lines of evidence summarized by Ganas et al. (2009) and Feng et al. (2010) indicate that the 2008 earthquake was associated with a 25 km long, sub-vertical right-lateral blind strike-slip fault, approximately 25 to 5 km deep. On the surface, most of the area is covered by an up to 3 km thick layer of flysch favoring plastic deformation.

The meizoseismal area of the 2008 earthquake is located at the boundary, an extensional province to the east, mainly represented by the Gulf of Corinth and a compressional province to the west, along the arc, mainly represented by the Ionian Islands, characterized by intense seismicity, compressional faulting and strong earthquakes (Bernard et al., 2006; Jackson et al., 1981; Mercier et al., 1979; Shaw and Jackson, 2010). The most recent major earthquake was the catastrophic 1953 Cephalonia earthquake sequence of magnitude 7.2 (Stiros et al., 1994).

These major earthquakes have produced coastal uplifts (Pirazzoli et al., 1996) and are assigned to thrusting, but focal mechanisms of relatively recent smaller (M<6.5) earthquakes tend to indicate that the west edge of the Aegean Arc is marked by a strike slip fault offshore Cephalonia (CTF; Fig. 1; Benetatos et al., 2004; Shaw and Jackson, 2010). There is evidence that strike slip faulting is important farther east as well: The 2008 earthquake seems to correlate with an about 150 km lineament, which may correspond to a segmented or continuous strike slip fault. To the SW this lineament corresponds to a series of smaller faults which ruptured during a period of about 20 years before 2008, but without any significant surface rupture, while to the NE, it corresponds to a fault delimiting the late Quaternary Gulf of Patras to the SE, regarded as a strike slip fault (RPF in Fig. 1; Chronis et al., 1991; Clews, 1989; Flotte et al., 2005; Piper et al., 1990; Roumelioti et al., 2004). Between this lineament and CTF other strike slip earthquakes have been found, indicating a >150 km-wide shear zone in the internal part of the arc (>200 km wide according to Floyd et al., 2010; Shaw and Jackson, 2010). Still, it is not possible to identify significant geologic traces of faults in the area because of the lithology, which is dominated by low-strength flysch deposits and especially by marine Pliocene and Quaternary sediments, uplifted up to several hundred meters (Piper et al., 1990; Stamatopoulos et al., 1998). This uplift increases to the east and northeast, and in the central part of the Gulf of Corinth its rate is probably more than 3 mm/year (see Fig. 4 in Stiros (1988) summarizing previous authors and Pirazzoli et al. (2004)). On the other hand, this 150 km long lineament is discordant to the E–W trending neotectonic lines in the area which indicate N–S extension (Lyberis, 1984) and to the high-order hydrographic network, which is of parallel type and constraint by the dominant tectonic trends (see Fig. 9 in Stiros, 1988).

Another limitation in the understanding of active tectonics is that seismicity in NW Peloponnese is poorly known, for in the 50–100 years it was affected by relatively small earthquakes (see below and Laigle et al., 2002), which may not be representative of the long-term situation, since at least for the city of Patras, there is evidence of destructive and conspicuously larger earthquakes in the last 200–2000 years (Stiros, 1995).

3. Deformation in the scale of 100 years

The analysis of the deformation of the area in a scale of approximately 100 years is based on historical triangulation data permitting estimates of the strain.

3.1. Data

Triangulation data analyzed come from the three historical surveys of the first-order triangulation network of Greece. The first first-order triangulation network was established and measured in 1890–1892, in the central and southern part of the country, and benchmarks consisted of pyramids. Measurements were made with much care and accuracy (60-period measurements) between stations in rather inaccessible mountain peaks, mostly >1500 m high, and with a mean spacing of 50 km. Summary measurements were published in the International Association of Geodesy (IAG) Proceedings, while the measurement specifications and their evaluation were reported by Stiros (1993). In 1929–1930 pyramids were replaced by pillars, in most places built at exactly the same position, and the whole network was remeasured with high accuracy (24-period surveys), partly in the framework of the international Project of Measurement of the Meridian Arc ("Struve Arc extension Project") administrated by IAG. Shortly after World War II the network was measured for a third time, between 1954 and 1965, but a large part of pillars, damaged during the war, was reconstructed, mostly in shifted positions.

Measurements of the three periods were of compatible, very high accuracy, as a result of the extreme care and of the high specifications with which the first-order historical surveys in Greece were made: the precision of the computed average values for angles was better than 0.00016^{g}, the misclosure errors in triangles were lower than 0.0002^{g}, and the dimensions of error ellipses in the three surveys were very similar (Stiros, 1993).

An analysis of the available data from the confidential archives of the Hellenic Military Geographic Survey (HMGS) revealed that a significant number of pillars of the early surveys had survived and were reoccupied during the three historical surveys. This permitted a first evaluation of the tectonic displacements of Greece based either on a comparison of the results of the three surveys (Stiros, 1993) or of the first, 19th century survey with GPS data in the 1990's (Biliris et al., 1991; Davies, et al., 1997). These studies, however, focused on a large area and on a priori assumptions of the stability of the length of certain baselines, permitted to compute displacements and contraction–extension.

In this article we focus on a small part of the overall network, covering the meizoseismal area of the 2008 earthquake. This network consists of 7 triangulation pillars surviving in the same position since their establishment in 1890/92 and common in the three surveys (Fig. 2; Table 1).

Observed angle differences were of the order of 10^{-4} to 10^{-5} degrees, in their majority statistically significant, because according to the Ferrero formula (Mikhail, 1976, p.88) for a maximum misclosure error of the order of 0.0002°, the upper bound of the uncertainty of the angles is 10^{-4} degrees.

Our analysis is based on the analysis of changes in measured angles, and this permits to estimate only the shear components of the deformation.

3.2. Methodology

If a rectangular is subject to shear, it will deform to a parallelogram, and its angles will change from their initial value of 90°. The theory of mechanics indicates that there is a relationship between the characteristics of deformation (defined by certain variables, usually regarded as tensor parameters) and the change (deformation) of angles. Assuming uniform deformation in the study area, for small changes of the angles, certain of the variables defining the deformation can be computed. This approach was adopted in this study.

The available triangulation data correspond to measured angles, each defined by three triangulation points. Changes in the value of specific angle between two surveys (of the order of 10^{-5} degrees in our case) permit to form an equation expressing the change of this angle as a function of variables defining the deformation. Repeated measurement of different angles (for instance of the angles of triangles) permits to form a system of equations, from which it is possible to compute the characteristics of the deformation, including the maximum shear and its direction, as well as the uncertainties of these estimates using least squares. The methodology is based on Bibby (1981), Prescott et al. (1979) and Savage (1983) and is summarized in Appendix 1.

In the case of observations of angles, only the maximum shear can be computed, and not the estimates of extension (which require observations of displacements), or of rotation (which require coordinate changes). Still, if several consistent observations of angle changes are available, an accurate estimation of the shear is possible. The output is rates of maximum shear $\dot{\gamma}$ and its direction Ψ.

3.3. Data analysis

The available data consist of a set of 15 angles measured in three surveys, covering the majority of the angles of the network shown in Fig. 2. Differences between consecutive surveys were of the order of 10^{-5} grad. Following the methodology described above and in Appendix 1, it was possible to compute shear for three intervals, I: 1890/1892–1929/1930, II: 1929/30–1950/1966 and I + II: 1890/1892–1950/1966, with an average spacing of 35 years between consecutive surveys. For the first two periods we used rates of angle change from simple differences of observations of directions. For the third period (whole survey interval) we computed rates of angle change using a simple regression analysis.

Our computations were based on the MATHEMATICA® and the MATLAB® software, and were made for the three intervals, and for different parts of the survey area:

(1) triangle 29–31–32 crossed by the 2008 seismic fault
(2) triangle 30–31–44 partly crossed by the 2008 fault and previous earthquakes
(3) polygon defined by areas (1) and (2)
(4) triangle 17–29–30 in the Patraikos Gulf area, to the west of triangle (1)
(5) triangle 29–32–31 just south of the city of Patras

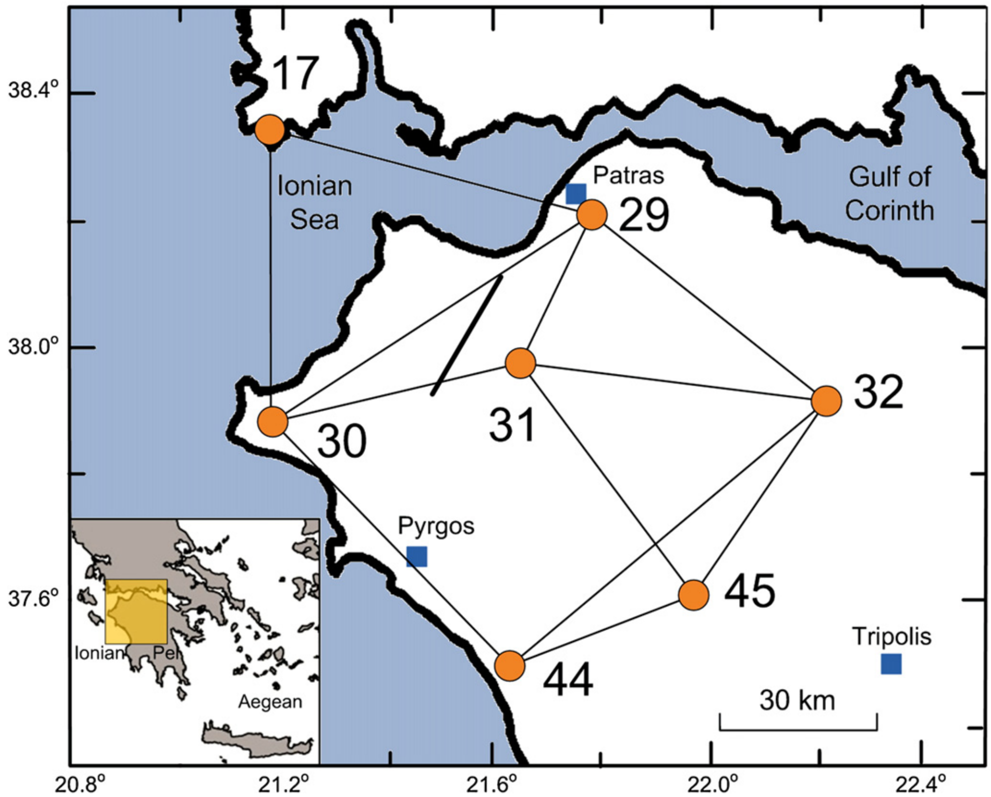

Fig. 2. The historical triangulation network examined — see Table 1 for additional info. Numbered dots indicate triangulation stations common in the three surveys, lines between the observed directions (triangulation observations) used in our analysis. A solid line indicates the predicted trace of the 2008 fault.

(6) polygon 29–32–45–32
(7) whole survey area.

These parts were selected on the basis of tectonic criteria and the availability of field data. The main results of our analysis are summarized in Table 2.

3.4. Evaluation of results

The above results are reliable for two reasons. First, because our study was limited to a small number of triangulation stations surviving in the study area in their initial positions. These stations were carefully built above a fixed underground benchmark in solid rocks, and most of the cases still survive. The possibility of misidentification of some of these benchmarks is totally excluded since only one single station existed in each high peak. In addition, pillars were inspected every about ten years, and the reports of their structural health permits to identify those which survived without problems during the whole study period, and in which our study was limited.

Second, the quality of historical triangulation surveys is assessed from the statistical analysis of observations in a chain of triangles implying strict geometrical conditions on measurements: sum of angles of 180° for each triangle in the chain (ignoring spherical excess for simplicity) and sum of 360° for angles around each station. Deviations from these values represent misclosure errors, which permit to estimate the accuracy (not simply precision) of the measurements, and recognize the statistical significance of the input in our calculations (see paragraph 3.1). The quality of data used is to a major degree due to the favorable landscape, measurements between mountain peaks and permitting minimum atmosphere-induced errors (Stiros, 1993). Triangulation measurements of high quality permit reliable estimates of shear, but as the theory of error propagation predicts (see Mikhail, 1976), even high accuracy observations of angles lead to estimates of displacements of low accuracy. A parallel is that modern accelerometer recordings are highly accurate, but the estimates of displacements deriving from their double numerical integration are not (Stiros, 2008). This is indeed a main reason why the value of historical triangulation data is usually underestimated or ignored.

A *first* conclusion from Table 2 is that the estimates of the maximum shear $\dot{\gamma}$ and Ψ of the first two periods I and II are less precise and consistent than the results of the whole period I + II, a result indeed expected on the grounds of statistics. For this reason we focus on the whole interval (I + II).

A *second* conclusion is that shear seems to be systematically more important to the west part of the area examined and decreases

Table 1
Summary of metadata of the Hellenic Military Geographic Survey for the triangulation stations examined in this article.

Old code	New code	Name	Est'd	Est'd in the same position	Est'd in a new position	Repairs	Visited/occupied
17	50350	Koutsiliaris	1892	1929	–	1983	1958/1959/1964/1977/1984
29	41030	Panachaiko	1891	1929	–	1952	1954/1964/1965/1983
30	40900	Chlemoutsi	1892	1929	–	1952	1954/1959/1964/1965/1977/1983
31	40940	Erymatnhos	1890	1929	–	1983	1954/1964/1965
32	40920	Kyllini (Zeria)	1890	–	–	1952	1930/1954/1965/1966/1983
44	40580	Minthi	1890	–	–	1952	1933/1936/1954/1964/1966
45	40680	Prof Elias (Levidi) Mainalon	1890	–	–	1952	1954/1963/1964/1965/1983

Table 2
Shear strain rates in various areas and time scales (intervals I, II, I+II). A statistically significant right lateral deformation increasing to the west is inferred. The direction of the maximum shear in the meizoseismal area of the 2008 and in the Gulf of Patras is similar to the strike of the 2008 fault. Bold data indicate stylistically significant results.

Area	Survey period	$\dot{\gamma}_1$ (μrad/year)	$\dot{\gamma}_2$ (μrad/year)	$\dot{\gamma}$ (μrad/year)	Ψ (°)
Triangle 29–31–30	I	−0.02 ± 0.01	0.01 ± 0.00	0.02 ± 0.01	29.9 ± 2.2
	II	0.64 ± 0.48	−0.28 ± 0.19	0.69 ± 0.49	33.0 ± 5.9
	I + II	0.30 ± 0.24	−0.14 ± 0.09	**0.34 ± 0.25**	**33.1 ± 6.1**
Triangle 30–31–44	I	0.01 ± 0.01	−0.00 ± 0.01	0.01 ± 0.01	44.5 ± 21.0
	II	−0.18 ± 0.08	−0.09 ± 0.06	0.21 ± 0.08	−31.8 ± 6.5
	I + II	−0.09 ± 0.03	−0.05 ± 0.02	**0.10 ± 0.03**	**−30.8 ± 5.7**
Polygon 29–31–44–30	I	0.01 ± 0.01	0.00 ± 0.01	0.01 ± 0.01	−44.2 ± 10.7
	II	−0.11 ± 0.13	−0.07 ± 0.08	0.13 ± 0.12	−28.5 ± 19.5
	I + II	−0.05 ± 0.06	−0.04 ± 0.04	0.06 ± 0.06	−27.0 ± 20.9
Triangle 17–29–30	I	−0.18 ± 0.18	0.16 ± 0.14	0.24 ± 0.17	23.5 ± 17.7
	II	−0.19 ± 0.13	−0.09 ± 0.10	0.21 ± 0.12	−32.7 ± 15.6
	I + II	−0.18 ± 0.02	0.04 ± 0.02	**0.19 ± 0.02**	**39.1 ± 2.8**
Triangle 29–32–31	I	0.02 ± 0.01	0.01 ± 0.01	0.02 ± 0.01	−29.4 ± 21.6
	II	0.22 ± 0.18	0.03 ± 0.21	0.22 ± 0.16	−41.5 ± 28.6
	I + II	0.12 ± 0.08	0.02 ± 0.10	**0.12 ± 0.08**	−40.8 ± 25.2
Polygon 29–32–45–31	I	0.02 ± 0.01	0.01 ± 0.01	0.02 ± 0.01	−29.6 ± 12.9
	II	0.18 ± 0.16	0.13 ± 0.16	0.22 ± 0.12	−27.3 ± 24.6
	I + II	0.10 ± 0.08	0.07 ± 0.08	**0.12 ± 0.06**	−27.5 ± 21.8
Whole network	I	−0.04 ± 0.05	0.05 ± 0.04	0.06 ± 0.04	19.0 ± 18.4
	II	0.01 ± 0.09	−0.01 ± 0.07	0.02 ± 0.09	33.0 ± 50.9
	I + II	−0.01 ± 0.05	0.02 ± 0.04	0.03 ± 0.04	15.1 ± 51.3

eastwards. This is evident from the fact that especially $\dot{\gamma}$ and Ψ become statistically more significant in the Gulf of Patras (triangle 17–29–30) and the western margin of the Peloponnese (triangles 29–31–30 and 30–31–44), in the 1-σ and even the 2-σ levels both concerning the rate of shear strain and its direction.

A *third* conclusion is that the shear in NW part of the study area is very much consistent with the 2008 co-seismic shear and expected in an area sandwiched between the 2008 and the Cephalonia transform fault (see Fig. 1); this is in particular evident for the direction of the max shear inferred from triangles 29–31–30 and 29–30–17 for which the angle of max shear was computed as 33.1° ± 6.1° and 39.1° ± 2.8°, respectively, little differing from the strike of the 2008 earthquake (30°, see Ganas et al., 2009).

A *fourth* conclusion is that there is a possibility for higher strain accumulation in the area around the 2008 fault. If this is true, then triangulation data have recorded strain built-up, partly at least released in 2008. The uncertainty level in the estimate of $\dot{\gamma}$ is high (0.34 ± 0.25), but the strain direction is clear (33.1° ± 6.1°). Relatively high strain is observed also in the Gulf of Patras (triangle 17–29–30).

A *fifth* conclusion is that the direction of the maximum shear in the Pyrgos area (covered by triangle 30–31–44) is about 60° different from those further north concerning the direction of strain (−31° compared to 30°). This perturbation of strain may be due to the relatively small earthquakes (Savage et al., 1981) which have hit this last area during the critical period (see Ganas et al., 2009 and below). Details on these earthquakes are certainly lacking.

3.5. Correlation with earthquakes

During this period only two significant earthquakes affected the wider area according to the catalog of Papazachos and Papazachou (1997). The 1925.07.06 earthquake of estimated magnitude 6.6 and approximate epicenter (Papazachos et al., 1997) SW of point 32 (φ = 37.8°, λ = 22.1°, close to station 32 in Fig. 2). It was an intermediate (~80 km) depth event which produced light damage in the Kalavryta (around station 32) and Tripolis area. The 1965.04.05 earthquake of magnitude 6.1 (epicenter φ = 37.4°, λ = 22.1°, to the south of the area shown in Fig. 2, depth ≤ 60 km) produced major damage in central Peloponnese (Papazachos et al., 1997).

According to Mc Kenzie (1972) and Ambraseys and Jackson (1990) the second earthquake was a much smaller event (M_s 5.9–6.0), with a different epicenter farther south (φ = 35.1°, λ = 24.3°, depth = 51 km), and had an essentially strike slip fault mechanism (125:74:−032). The major damage (collapse of about 7% of the 100,000 houses in the wider area) and death toll (18 people killed) was assigned to the poor construction style of the houses during this period (remembered by one of us, SCS) and their unstable foundations, with 70% of damage associated with local ground instability in high gradient terrain in flysch and shales. The numerous strong aftershocks were also assumed to have contributed in the high damage toll. Both these earthquakes are unlikely to have caused significant surface deformation in the area covered by the triangulation network examined.

Apart from these earthquakes, Ganas et al. (2009) summarize a number of other smaller and rather poorly known events which have affected the wider Pyrgos area. As noticed above, these earthquakes may be responsible for the perturbation of the strain in that area.

3.6. Correlation with GPS

The historical triangulation data were compared with results of GPS surveys permitting estimates of displacement vectors for a period of about 10 years (Fig. 3; Floyd et al., 2010; Hollenstein et al., 2008). The location of the GPS stations in the study area is compatible with those of historical triangulation data, and permits comparison of the longer and shorter term estimates of shear. Estimates of the order of 0.2 μrad/year (30 mm/150 km) can be deduced from Fig. 15 in Floyd et al. (2010), which are compatible with those shown in Table 2.

The overall conclusion is that shear deduced from GPS data covering a period of 10 years before the 2008 earthquake is compatible with that derived from historical triangulation data covering a period of approximately 70 years.

4. Deformation in the scale of 10^5–10^6 years

4.1. Hydrographic network and strike slip faulting

Strike slip displacements along faults tend to produce offset streams (Sylvester, 1988). If the relief is young, such stream offsets also permit to characterize and date the deformation (Ferry et al., 2007; McCalpin et al., 2009; Yeats et al., 1997). The relief in NW Peloponnese offers the challenge to identify possible geologically recent strike slip displacements: a large part of the study area is

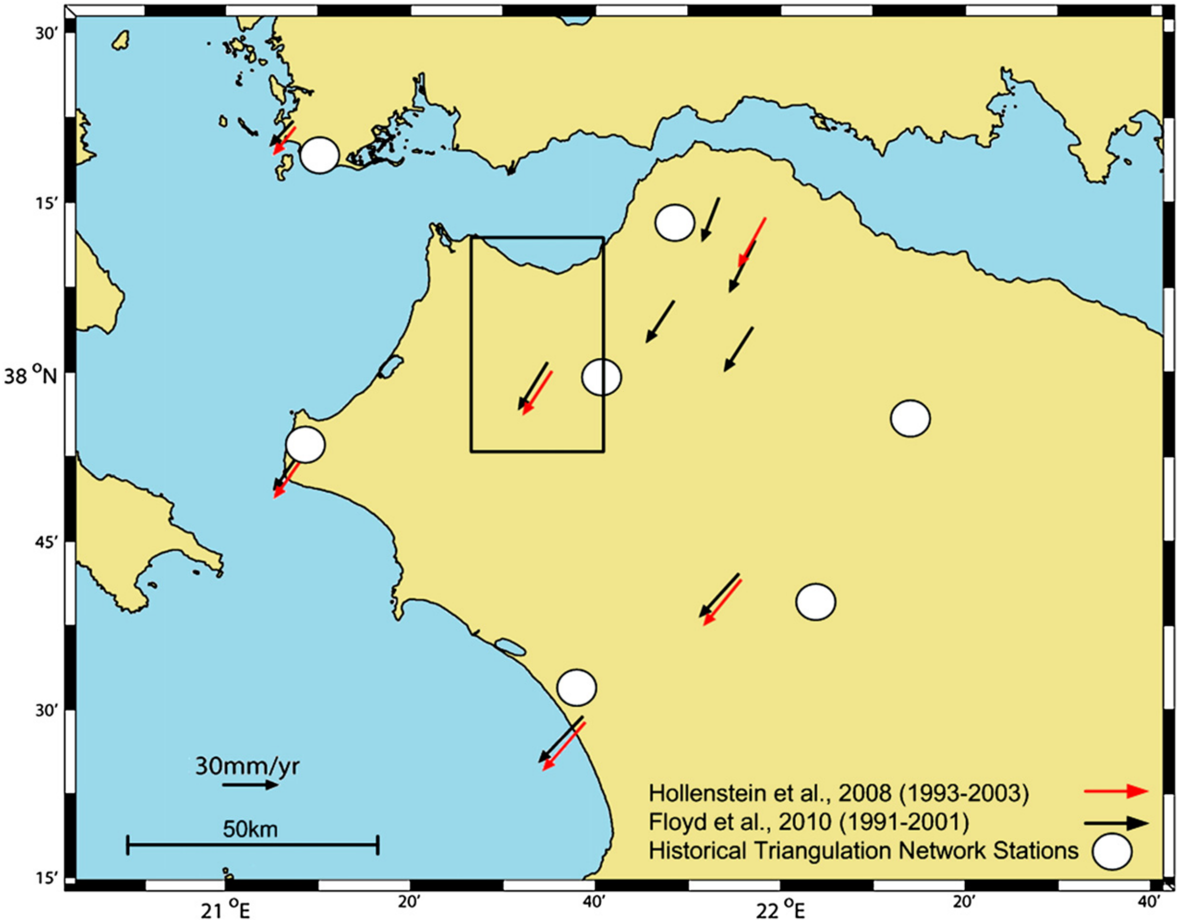

Fig. 3. GPS displacement field in the study area for a period of approximately 10 years. Black arrows after Floyd et al. (2010), red after Hollenstein et al. (2008). White circles indicate the triangulation stations of Fig. 2.

covered by Pleistocene and Pliocene deposits (Bornovas and Rondogianni, 1983), uplifted up to several hundred meters (see Fig. 4 in Stiros, 1988) and hence represents a "young" relief, on which the important geologically recent morphotectonic events have left their signature.

This approach is new for the region, because of the dominance of normal faulting, morphotectonic analyses of offsets of the hydrographic network are extremely rare. Still, a small-scale study covering the whole of North Peloponnese provided evidence of a hydrographic network favorable for the identification of strike slip offsets in the vicinity of the trace of the 2008 fault (cf. Fig. 9 in Stiros, 1988). This prediction was verified by a preliminary analysis of 1:50,000 scale topographic maps and encouraged a further study.

4.2. Data and results

Maps of two scales, 1:50,000 and 1:5000 of the Military Geographic Survey are available in Greece. The 1:5000 scale map is too detailed to study a 25-km long fault. The smaller, 1:50,000 scale map permits an about 15–20 m threshold for detection of systematic offsets in map data (contours, streams, etc.) and was selected for this study.

The hydrographic network of NW Peloponnese around the predicted surface trace of the 2008 fault was compiled and is shown in Fig. 4. The hydrographic network in the critical area is highly developed and mostly of dendritic type, so that streams have a nearly random distribution concerning the direction of their segments and the chance for a surface fault to cross streams of different orders is high. In addition, the area of Fig. 3 covers different geologic formations, hence systematic effects related to a specific lithology (amplification or attenuation of a certain type of deformation) are not expected.

A detailed examination of Fig. 4, as well as of the maps from which this figure was produced, provides no evidence of stream offsets both along the predicted surface trace of the 2008 earthquake and the wider area covered by this Figure.

It can therefore be concluded that the hydrographic network provides no sign of strike slip deformation in the wider zone of the predicted trace of the 2008 fault, either as a single or a segmented fault.

4.3. Significance of the absence of offset streams

The absence of signs of offset stream in the young relief of the study area can be indicative (a) of an absence of faulting above the threshold of identification of offsets for the selected map, (b) of an analysis in a map scale too small to permit identification of stream offsets-strike slip movements or (c) of faulting non-preserved because of various reasons. In order to analyze points (b) and (c), the hydrological conditions of the study area were examined.

The study area is characterized by relatively high precipitation rates (mean precipitation > 800 mm/year; Hatzianastassiou et al., 2008) to a major degree associated with storms (rainfalls of the order of 150 mm/24 h; Koutsoyiannis and Baloutsos, 2000; Koutsoyiannis, unpublished data). Under these conditions small-scale offsets produced by faulting are expected to be rapidly obliterated. An example of such transient

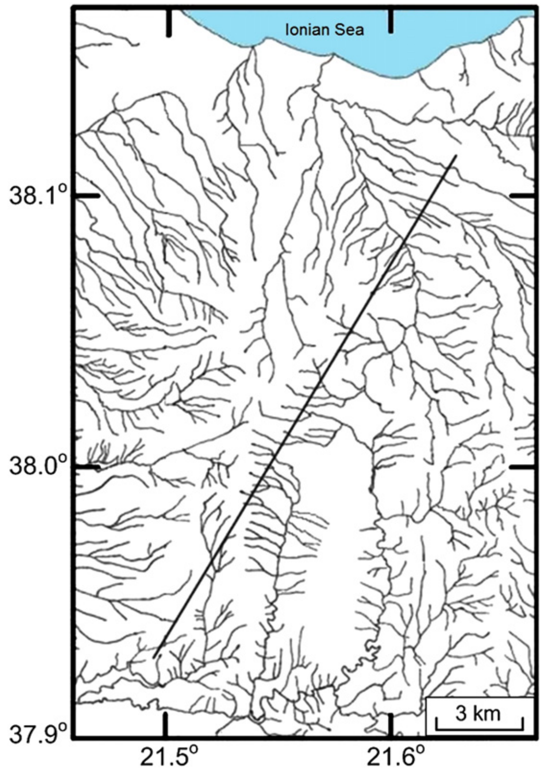

Fig. 4. Hydrographic network in the vicinity of the predicted trace of the 2008 fault (solid line), compiled from 1:50,000 scale topographic maps of the Military Geographic Survey of Greece. No evidence of steam offsets, indicative of strike slip faulting along the inferred fault-trace and in the wider area exists. For location see rectangle in Fig. 3.

changes in the course of a stream, not away from the study area, at a scale of <100 years has been indeed presented by Zelilidis (2000). This indicates that a study in a finer map scale would not produce additional substantial evidence, as would be the case in arid or semi-arid environments (Ferry et al., 2007; Yeats et al., 1997).

On the other hand, although no studies of stream offsets by strike slip faulting have been made in the past in the study area and the wider region, there has been clear evidence of stream offsets in normal faulting environments in Northern Peloponnese, by secondary (transfer) faults (Doutsos and Piper, 1990). Such offsets are clearly visible in 1:50,000 maps. For this reason, it can be concluded that Fig. 4 indicates absence of significant faulting prior to 2008.

5. Discussion

New evidence presented above permits to constrain the deformation in the vicinity of the 2008 fault, the first major strike-slip fault inferred to have reactivated in Greece, in different time scales. At a time scale of 10^5–10^6 years, morphotectonic data provide no evidence of significant fault activity, since no signs of offsets in the hydrographic network in a new landscape have been found. At the scale of 10^1–10^2 years, historical and modern geodetic data provide evidence of (aseismic) shear consistent with the fault kinematics derived from seismological data. These results may indicate that the strike slip deformation is new; perhaps the strike slip earthquakes summarized in Fig. 1 are the first, or among the first to have occurred in the area. This justifies the hypothesis of Feng et al. (2010) for an immature fault.

Still, the correlation of the 2008 fault with a lineament of other faults to the SW which reactivated as strike slip faults during the last 20 years and with the RPF, regarded also as a strike slip fault to the NE (Fig. 1; Clews, 1989) may indicate a pre-existing dormant or fossil fault which has recently (in the last 20 years?) entered a new phase of activity.

There are also several other points to notice.

First, as can be deduced from Fig. 1 and the data presented above, the deformation in the west edge of the Aegean Arc is accommodated by an about 150 km wide shear zone along the internal part of the arc, between the Cephalonia fault offshore and the 2008 fault (see also Floyd et al., 2010; Shaw and Jackson, 2010). East of the 2008 fault shear becomes smaller and diffuse (see Table 2). An analogy with the seismic deformation at Sumatra (Satriano et al., 2012), with pre-existing crustal intra-plate discontinuities reactivated as strike slip faults, may be noticed.

Second, the apparent gradual rupture of the a major fault, 100–150 km long, is somewhat reminiscent of the gradual rupturing of the North Anatolian Fault (Atakan et al., 2002; Stein et al., 1997; Toksoz et al., 1979). If this is true, then reactivation of RPF, marking the east margin of the Late Quaternary Gulf of Patras (Fig. 1; Chronis et al., 1991; Piper et al., 1990) may have certain important implications for the seismic risk for the city of Patras.

Third, the identification of aseismic deformation in the study area is expected, for aseismic deformation seems to dominate large parts of Greece (Jackson, 1994; Laigle et al., 2002).

Fourth, the preseismic surface shear seems to contrast with the absence of significant co-seismic surface deformation. A possible explanation is that the upward propagation of the seismic displacements is controlled by the level of strain, with the uppermost layer of weak rocks acting as a low-pass filter: in low strain rates (interseismic period) shear reaches the surface, but not in high strain rates (2008 earthquake); a parallel is the mechanism to avoid slamming doors; doors open only when smoothly pushed. Whether this behavior is characteristic of the period of fault immaturity (Fialko et al., 2005), and of other faults as well, is an open question.

A remaining problem is the relationship between shear and thrusting along the west part of the Aegean Arc. Focal mechanisms of earthquakes in the last 20 years tend to indicate strike slip in the zone between the 2008 fault and the CTF (Fig. 1; see also Floyd et al., 2010; Shaw and Jackson, 2010). Major earthquakes, on the contrary, such as the Cephalonia 1953 sequence of magnitude 7.2 was associated with coastal uplift and thrusting (Stiros et al., 1994), in accordance with structural data (Mercier et al., 1979; Underhill, 1989).

This interplay between thrust and strike-slip faulting is beyond the scope of this paper, and may reflect an effect of broader scale (Kreemer and Chamot-Rooke, 2004), because a similar situation seems to exist in the eastern edge of the Aegean Arc, in the Rhodes Island: recent moderate earthquakes indicate an essentially strike slip margin, the island is dominated by normal faulting, while raised beaches and the overall tectonic deformation tend to indicate thrusting (Benetatos et al., 2004; Kontogianni et al., 2002; Shaw and Jackson, 2010; Yolsal-Çevikbilen and Taymaz, 2012).

6. Conclusion

A study of the deformation of the meizoseismal area of the 2008 NW Peloponnese earthquake, of the first important strike slip fault identified in continental Greece, was made for two different time scales. A possible conclusion is that the activity of this fault started only recently, reflecting an immature or even a recently reactivated dormant fault, discordant to the young relief. While this fault was not associated with significant surface seismic deformation, it is characterized by interseismic shear, indicating that the uppermost flysch layers may behave as filters to high-rate seismic strain. Successive reactivation of segments of this fault may provide some constraints for seismic risk in the area, while the relationship interplay between an inferred zone of shear along the arc and thrust faulting seems unclear.

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Appendix 1

It is assumed that the deformation of the area is uniform and planar (depending only on the orientation/azimuth), hence it and can be described by the symmetrical tensor

$$T = \begin{pmatrix} \varepsilon_{11} & \frac{\varepsilon_{12}}{2} \\ \frac{\varepsilon_{12}}{2} & \varepsilon_{22} \end{pmatrix}. \quad (1)$$

Let us assume that during a certain survey, k, three triangulation stations describe an angle $\varphi_{ij,k}$ as the difference of the azimuths α_i, α_j of two lines (i and j taken in clockwise direction), so that

$$\varphi_{ij,k} = \alpha_{jk} - \alpha_{ik}.$$

The same measurements were repeated during a subsequent survey $(k+1)$, so that

$$\varphi_{ij,(k+1)} = \alpha_{j(k+1)} - \alpha_{i(k+1)}$$

and permit to compute the (observed) angle difference

$$\delta\varphi_{ij,k} = \varphi_{ij,(k+1)} - \varphi_{ij,k} \quad (2)$$

between the two surveys $(k+1)$ and k. This difference is regarded as a function of the shear deformation affecting the study area.

Since the changes in the azimuths between the two surveys were infinitesimal (of the order of 10^{-5} grads, see Section 3.3), the observed angle difference $\delta\varphi_{ij,k}$ can be expressed as a function of the (approximate) values of azimuths i and j and of γ_1 and γ_2

$$\delta\varphi_{ij,k} = \gamma_1\left(\sin 2\alpha_j - \sin 2\alpha_i\right)/2 + \gamma_2\left(\cos 2\alpha_j - \cos 2\alpha_i\right)/2 \quad (3)$$

with

$$\gamma_1 = \varepsilon_{11} - \varepsilon_{22} \text{ and } \gamma_2 = 2\varepsilon_{12} \quad (4)$$

reflecting the shear strain components in a two-dimensional Cartesian coordinate system.

γ_1 measures the right-lateral shear across a vertical plane striking N45°W (or left-lateral shear across a vertical plane striking N45°E), and γ_2 the measures right-lateral shear across a plane striking eastwards (or left-lateral shear across a vertical plane striking northward) (Savage, 1983).

If instead of angle changes, (annual) rates of angle changes $\delta\dot{\phi}_{ij,k}$ are computed,

$$\dot{\gamma}_1 = \dot{\varepsilon}_{11} - \dot{\varepsilon}_{12} \text{ and } \dot{\gamma}_2 = 2\dot{\varepsilon}_{12} \quad (5)$$

Eq. (3) is replaced by Eq. (6) connecting (annual) shear rates with rates of angle change

$$\delta\dot{j}_{ij,k} = \dot{\gamma}_1\left(\sin 2\alpha_j - \sin 2\alpha_i\right)/2 + \dot{\gamma}_2\left(\cos 2\alpha_j - \cos 2\alpha_i\right)/2. \quad (6)$$

From $\dot{\gamma}_1$ and $\dot{\gamma}_2$ the value of maximum shear $\dot{\gamma}$ and of its azimuth ψ can be computed using the formulae

$$\dot{\gamma} = \left(\dot{\gamma}_1^2 + \dot{\gamma}_2^2\right)^{1/2} \quad (7)$$

$$\Psi = 0.5 \arctan(-\dot{\gamma}_1/\dot{\gamma}_2). \quad (8)$$

Estimates of $\dot{\gamma}$ are dimensionless, they are expressed in terms of radians, and in tectonically active areas their usual values are of the order of μrad/year and describe the change of a right angle (90°) at the direction of the maximum shear (Prescott et al., 1979; Savage, 1983).

Assuming planar and symmetric deformation (Bibby, 1981), for each measurement of angle change rate between two consecutive surveys one equation similar to (6) can be written. For a triangle with measured three angles in the two surveys, three equations can be formed, and permit to solve a redundant system of three equations with two unknowns. For a wider area covered by a triangulation network such as that of Fig. 2, each measured angle contributes to one equation and leads to a redundant system of equations with two unknowns. The solution of this equation using conventional least squares techniques permits to compute the best estimates of $\dot{\gamma}_1$ and $\dot{\gamma}_2$, as well as their variance–covariance matrix Σ, a 2×2 symmetrical matrix. The values of the maximum shear rate $\dot{\gamma}$ and of the maximum shear direction angle (Ψ) and of their standard deviations were subsequently computed using Eqs. (7) and (8) and the law of error propagation.

$$\Sigma_{\dot{\gamma}} = J_{\dot{\gamma}} \cdot \Sigma \cdot J_{\dot{\gamma}}^T \qquad \Sigma_{\Psi} = J_{\Psi} \cdot \Sigma \cdot J_{\Psi}^T \quad (9)$$

in which $J_{\dot{\gamma}} = \left[\frac{\partial\dot{\gamma}}{\partial\dot{\gamma}_1}\ \frac{\partial\dot{\gamma}}{\partial\dot{\gamma}_2}\right]$ and $J_{\Psi} = \left[\frac{\partial\Psi}{\partial\dot{\gamma}_1}\ \frac{\partial\Psi}{\partial\dot{\gamma}_2}\right]$ are the Jacobians of Eqs. (7) and (8), respectively.

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