COULOMB STRESS CHANGES IN THE GULF OF CORINTH (GREECE) FOR THE 1992-1995 PERIOD

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ABSTRACT

The rift of the Corinth Gulf, one of the most active areas, has attracted attention for more than a decade and been extensively monitored by geodetic, seismological and various geological observations in order to evaluate its complicated tectonic behaviour.
Recent studies have verified that the rift is bordered on both its north and south sides by active faults, predominantly normal ones. However, much of the seismicity in the region is attributed to off-shore active faults. Their geometry and tectonic activity is estimated indirectly via analysis of the seismic episodes, geodetic observations on the free surface, observations of microseismicity in the area, multibeam bathymetry etc.
Although the 1995 Aigion earthquake, a well documented event resulting from normal faulting, has been most rigorously studied, the preceding Galaxidi-Itea earthquake of M_s~5.9, of November 1992, beyond the early studies that estimated its characteristics, has not merited a similarly prolonged and detailed analysis, possibly due to its minor consequences in the vicinity.

The present work deals with the assessment of the Coulomb stress change (ΔCFF) associated with the Galaxidi-Itea 1992 earthquake, assuming the earthquake can be modeled as a static dislocation in an elastic half-space. The stress changes are estimated for normal faulting and for the appropriate parameters of the Galaxidi event. It is, also, investigated whether this earthquake may be considered as a precursor for the 1995 Aigion earthquake of M_s~6.2.

1. INTRODUCTION

Normal faulting, in actively extending regions of the continents, is usually organized into sub-parallel systems distributed over areas of tens or even hundred kilometres wide. In Greece, one of the most prominent and active features of such a system of faults is the rift along the Gulf of Corinth.
The Gulf of Corinth is the most rapidly extending rift system in Greece with about 120km length and 30km width and a WNW-ESE trend. It is believed to be active at the present rates since the last 5Myrs.It is connected with one of the highest seismic activities in the Euro-Mediterranean region: 5 earthquakes of magnitude greater than 5.8 in the last 35 years, an estimate of 1-1.5cm/yr of north-south extension, frequent seismic swarms, and destructive historical earthquakes [7, 11, 20]. The rift appears to be asymmetric and bounded along its south coast, on Peloponnesos, by north-dipping faults of en échelon pattern with maximum segment lengths of 15-25km.

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Seismic, geomorphologic and geological observations suggest that the Peloponnesos coastal and offshore normal faults seem to be the most active today, resulting in the long term subsidence of the northern coast, and on the upward displacement of the main footwalls. These vertical motions relatively to sea level are superimposed on the general uplift of the northern Peloponnesos [5].
It is estimated that for the central part of the rift all recent large earthquakes (Eratine of Phokida, M=6.3, 1965; Antikyra, M=6.2, 1970; Galaxidi, M=5.8, 1992, Aigion, M=6.2, 1995) activated offshore faults with shallow north-dipping planes [2]. At least for the 1995 and, possibly, for the 1992 events [4, 10] these planes are shown to be the fault planes. These 30° to 35° dip angles differ significantly from the steeper 45° to 50° dip angles of the eastern part of the rift, such as the ones for the three earthquakes of the Corinth sequence of 1981 (Figure 1) [11].
Present day estimates, derived from GPS observations, suggest fault slip rates of the order of 10-15mm/yr for the western part of the Gulf, not quite in agreement with observed uplift rates of 1-2mm/yr over the last 0.3Myrs [1, 15]. However, the extension rates estimated from GPS observations, for the eastern part of the Gulf, are significantly lower (of the order of 5-6mm/yr) [7].

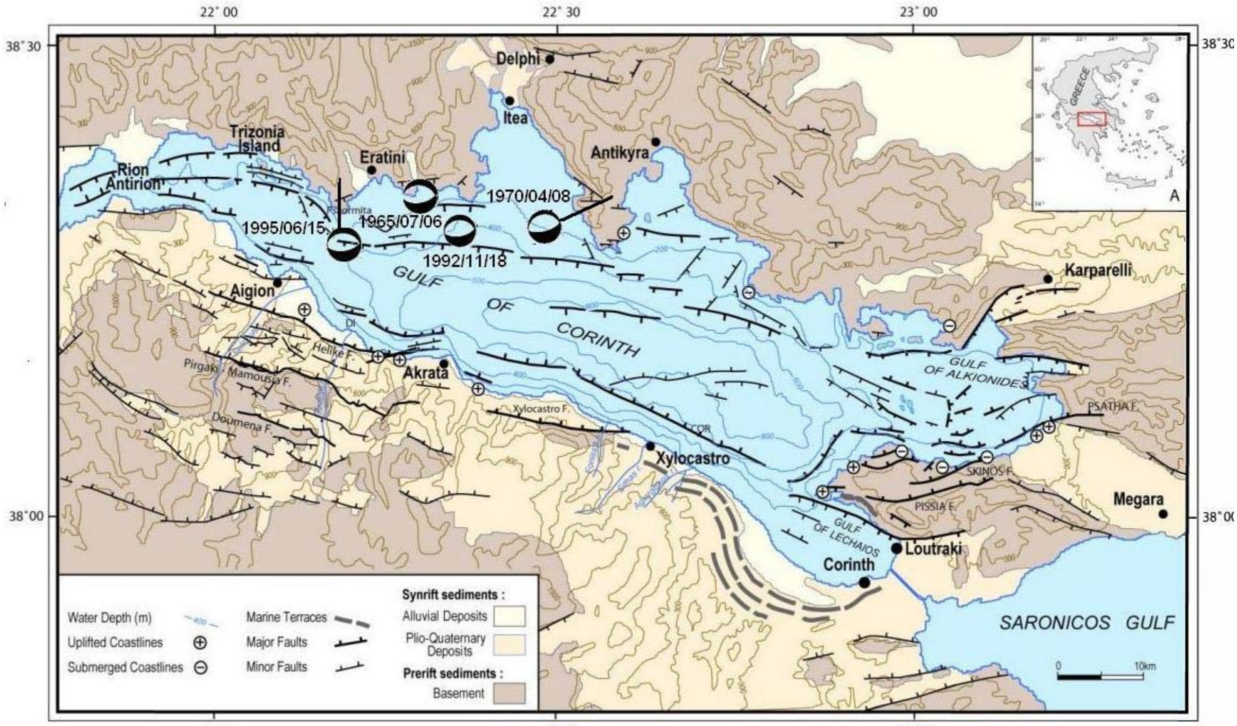


Figure 1 The geological map of the Corinth Gulf with the recent major earthquakes for the western part. Modified from [14, 17].

2. STATIC STRESS CHANGE

Stresses around an active fault are accumulated slowly due to the lithospheric plates´ motion. If the stresses surpass the strength of the crust a fracture takes place which causes a seismic event. Thus, part of the accumulated stresses is released on this fault while the stress field in neighbouring areas is changed. A measure of this change is the so-called Coulomb stress, which is the difference between the shear stress in the fault direction and the shear strength, assuming that the Mohr-Coulomb criterion expresses the strength of the crust material.

Therefore, if the distribution of the static stress changes around the ruptured fault is known it may provide useful information whether the seismic activity on a nearby fault will accelerate or not depending on the increase or decrease of Coulomb stress induced by the ruptured fault.
Several studies have indicated that relatively small but sudden changes of the stress field applied on the faults may affect the rate of seismic activity in the surrounding area [13, 23, 25]. Such stress changes, of the order of 0.2MPa (≈2 bar) are only a fraction of the stress drop during an earthquake and were regarded insignificant, until a few year ago. A reasonable question would be how such a small stress increase, equal to the pressure at the base of 10 m high embankment, may advance a rupture. This may be evidence that the faults in an active area are in a limit state, very close to fracture or slipping; thus a small change could trigger the onset of fracture.
Although the stress distribution inside the lithosphere is unknown the stress change due to a rupture on a fault may be estimated. Therefore, the Coulomb stress is calculated considering the lithosphere as a homogeneous elastic, isotropic half space, an assumption that simulates the crust behaviour satisfactorily. Elastic dislocations on rectangular planes -representing the faults- in this half space are used [18] and the calculations require the knowledge of both the fault geometry and the neighbouring stress field.
Thus, the Coulomb stress change, that is the difference of the absolute value of the shear stress developed in a particular direction minus the shear strength along it, is given by the following expression:

(*1*)

where:
$\Delta\tau_{slip}$ is the change in shear stress (positive along fault slip direction) due to the first earthquake resolved in the slip direction of a second fault, $\Delta\sigma_n'$ is the normal stress change (positive if compressive) due to the first earthquake, resolved in the direction orthogonal to a second fault plane and φ is the friction angle of the fault surfaces. ΔCFF is resolved onto the fault plane and in the slip direction of a second "receiver" fault, and at the hypocentre of the second fault.
If ΔCFF is positive, that is the shear stress on the fault exceeds its shear strength, then the rupture is accelerated. In this case the first earthquake may bring the second fault closer to failure. While, if ΔCFF is negative, the normal stress increase augments the shear strength. In this case the first event may send the second event farther away from failure, and into a so-called stress shadow. The stress shadow lasts as long as it takes the second fault plane to recover from the stress decrement. One manner of recovery is through long-term tectonic loading.
The method presumes that the static stress drop due to a large earthquake is recovered during the time interval that elapses until a new major event takes place; the static stress change in a seismic cycle is zero. Furthermore, it is often assumed that the initial Coulomb stress in the area of interest is zero. Although this may not be true it does not affect the computations, since the element of interest, in the vicinity of an earthquake, is not the static stress but its change. However, it should be acknowledged that the stress field of an area is the combination of the long-term tectonic loading due to the lithospheric plates′ motion and of the seismic ruptures. The size and orientation of the stress field due to seismic rupture may be estimated by studying the focal mechanism solutions of the seismic events occurring in the vicinity. But a rather prolonged time period of observations is needed in order to acquire a reliable estimate of the orientation of the stress field. Even then, the components of the stress field tensor are only known in specific locations, those of the explicit seismic events and do not represent the overall stress filed. Geologic observations in situ are necessary in order to estimate the local stress field [24]. A simpler approach is to use empirical formulae in order to estimate the stress drop accompanying a fault rupture for the neighbouring area.

3. DATA ANALYSIS

In 1992, November the 18th, an earthquake of magnitude M_S = 5.9 occurred in the central part of the Gulf of Corinth. No surface faulting was observed and the event was mainly felt in the towns of Itea and Galaxidi, where all the damage has been reported [6, 12]. These observations indicate an offshore location for the fault that gave this earthquake, probably close to the northern border of the Gulf. The location of the main shock was rather controversial due to lack of close by seismic stations. However most of the research centres, such as USGS and Thessaloniki Observatory, located the event offshore closer to the northern border of the Gulf and in front of the Xylocastro fault, one of the prominent active normal faults [6] (Figure 1).

Although the depth was rather well constrained, at 7.4±1 km, and compatible with the depth distribution of micro-earthquakes recorder in the end of 1991[20], the fault parameters (Table **2**) are less well so due to the poor azimuthal coverage of stations toward the southwest [10]. Surprisingly, this earthquake was not followed by a strong aftershock sequence and no aftershocks of magnitude greater than 3.1 were observed during the days after the main event. According to[10] this event, as well as the one in 1965, close to Eratini (Figure **1**), took place between two important active normal faults, the Helike and Xylokastro, that are offset at the surface by about 5km and presumably remain so at depth. This offset could behave as a barrier possessing greater strength than the surrounding two faults. But, since the barrier is small and the extension rate across the Gulf is larger than 1cm/yr, it has to evolve over time as an asperity and break without generating aftershocks as the Helike and Xylokastro faults are weaker.

The distribution at depth of the limited sequence of aftershocks associated with this event shows a dip angle of 30° toward the north in accordance with the dip angle of the fault plane deduced from the focal mechanism [10].

In the present work an assessment of the Coulomb stress change (ΔCFF) associated with the 1992 Galaxidi earthquake and its impact on the stress field of the surrounding area, especially on the offshore fault of the 1995 Aigion earthquake and the closer active onshore faults has been carried out.

The software used in this analysis was the Coulomb software, version 2.6 [25]. It calculates, on any surface and at any depth, static displacements, strains and stresses caused by fault slip, point sources of inflation/deflation, and dike expansion/contraction. All calculations are made in a half-space with uniform isotropic elastic properties. The software implements elastic dislocation formulae [18] and boundary element formulae [8].

After a fault rupture occurs, the stress changes on the specified ("receiver") fault are determined as function of the direction and magnitude of the fault slip and the regional in situ stress field. The programme may, also, calculate Coulomb stress changes on planes with optimum (critical) orientation for a comparison with the actual specified faults. It is presumed that a sufficient number of fractures (small faults) exist having all possible orientations and that the faults optimally oriented for failure will be the most likely to slip in small earthquakes. It is important to emphasize that the change of Coulomb stress function is calculated on the specified orientations, as well as, on planes of optimal orientation with respect to the regional stress field. [13].

The elastic parameters necessary for the model were either taken from the USGS (NEIC) data or were computed (Table 1). Thus the shear modulus (or modulus of rigidity) was computed as:

(2)

where E is the Young modulus and ν is the Poisson ratio.
To estimate the stress drop ($\Delta\sigma$ in bars), the following expression was used [3]:

(3)

where μ is the modulus of rigidity and *Mo* is the seismic moment expressed in (dyn·cm).
The stress changes were estimated for normal faulting and for the appropriate parameters of the geometry and slip direction of the 1992 Galaxidi earthquake rupture. It should be mentioned here that, for stress change estimations, the exact details of the geometry and slip of a main event become less important the farther one goes from the rupture. The fault parameters of both the 1992 and 1995 events (Table 2) were chosen from [6, 7].
The choice of the appropriate "receiver" faults was based on studies carried out on recent micro-seismicity (2000-2001) and other geophysical data for the western and central parts of the rift. Among the active, almost parallel, faults of this part of the Gulf (Figures 1, 2a, 2b) those that might be affected by the stress change due to the 1992 and 1995 events and with known geometries were chosen as "receiver" faults (Table 3). The geometry parameters were taken from: http://www.ingv.it [19]. The parameters for the Xylokastro fault [22] are less well established since it is not so thoroughly studied as the Helike and Aigion ones of the western part. Its depth, dip and rake are chosen to be compatible with the other two as they appear to belong to a common fault system [15] although separated by the structural culmination of Zarouchla which acts as a barrier interrupting the lateral propagation of these faults [9]. Therefore, it was chosen not to be modeled as a dislocation in the Coulomb analysis like the rest, but to examine whether it was affected by the 1992 earthquake.

Model Parameters				
Young modulus *(GPa)*	**Shear modulus** *(GPa)*	**Poisson' s ratio**	**Coefficient of friction**	**Calculation depth** *(km)*
73.2	29	0.28	0.6	7

Table 1 Model elastic parameters

1992 and 1995 Earthquakes										
Date	**Ms**	**Mo** *(Nm)*	**Lat** *(o)*	**Long** *(o)*	**Length** *(km)*	**Strike** *(o)*	**Dip** *(o)*	**Rake** *(o)*	**Top/Bottom** *(km)*	**Slip** *(m)*
18/11/1992	5.9	0.5×10^{18}	38.30	22.45	14	270	30	-81	5.2/9.7	0.21
15/06/1995	6.2	3.9×10^{18}	38.36	22.20	15	277	35	-81	4.5/9.7	0.87

Table 2 Parameters for the offshore faults of the 1992 and 1995 earthquakes [*6, 7*]

The principal stresses of the "regional" stress field for the 1995 event were chosen as: σ_1 = 27bars, equal to the estimated stress drop due to the 1995 event, σ_3=0 and the intermediate σ_2 = 13.5bars, while the principal axes were taken from USGS (NEIC) [16]. The respective stress drop for the 1992 event was found as: σ_1 = 72bars, σ_3=0 and the intermediate σ_2 = 36bars while their orientation was taken, as averaged from fault mechanism solutions for the whole of the Gulf from [21] (Table 4).

Fault Parameters							
Fault	**Lat** *(o)*	**Long** *(o)*	**Length** *(km)*	**Strike** *(o)*	**Dip** *(o)*	**Rake** *(o)*	**Top / Bottom** *(km)*
East Helike	38.193	22.150	16.6	279	50	270	0.2/7.5
West Helike	38.231	22.030	12	283	50	270	0.2/7.5
Aigion	38.265	22.035	10	277	50	270	0.2/7.5
Xylokastro*	38.10	22.50	15	280	50	270	0.2/7.5

Table 3 Fault geometry for "receiver" faults [19]. *The Xylokastro data are approximated from [22] and the fault has not been modeled as a dislocation in the Coulomb analysis.

Principal Axes						
	σ_1		σ_2		σ_3	
	Azimuth *(o)*	**Plunge** *(o)*	**Azimuth** *(o)*	**Plunge** *(o)*	**Azimuth** *(o)*	**Plunge** *(o)*
1992	272	72	92	18	2	0
1995	174	12	82	9	316	75

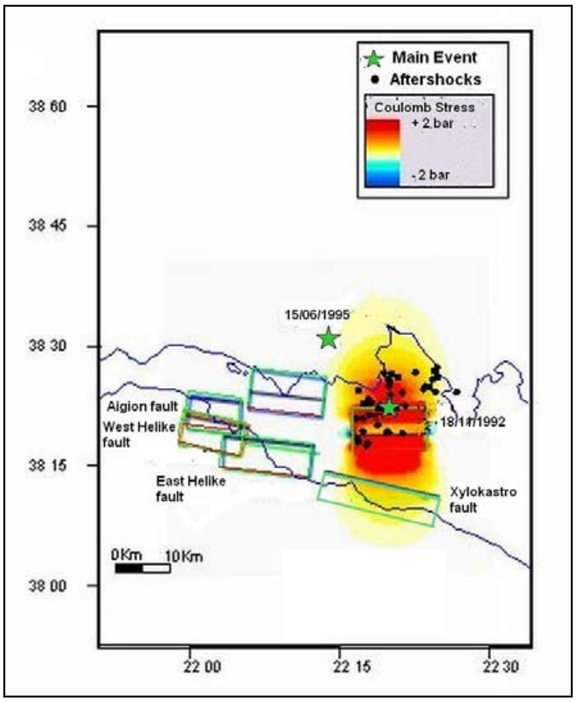
Table 4 "Regional" in-situ stress field orientation for 1992 and 1995 events [21, 16].

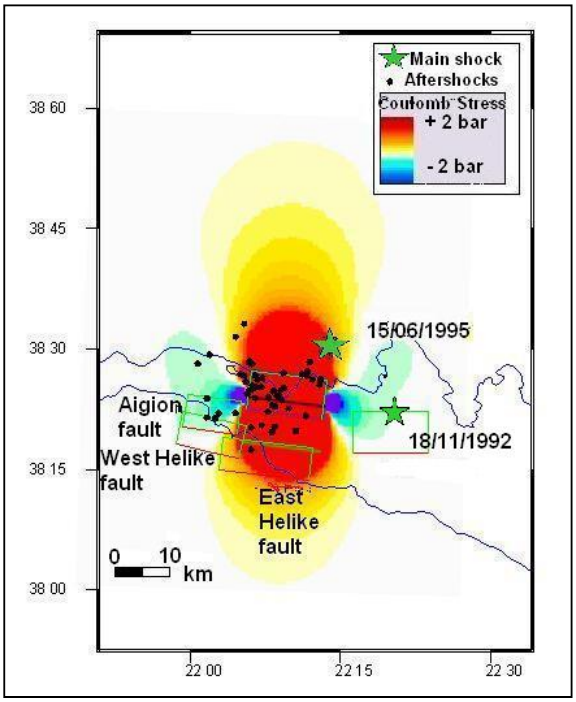
All calculations of ΔCFF refer to a depth of 7km. This depth was chosen as being halfway between the top and bottom depths of the fault planes, as well as the middle of the seismogenic zone in the area. Cross-sections for the 1992 event are depicted to a maximum depth of 25km below the free surface, while for the 1995 down to 40km. The average orientation of the most active well documented faults on the north coast of Peloponnesos (Table 3) was also chosen as a specified orientation.
The distribution of the 1992 aftershock sequence was compared with the positive ΔCFF change area (Figure 2a, 2b). In order to compare the behaviour of the two major consecutive earthquakes the respective map view and cross section for the 1995 event together with the aftershock distribution are also depicted [16] (Figure 2c, 2d).
Finally, the combined Coulomb stress change for 1992 and 1995 is depicted together with their respective aftershock sequences in order to examine whether an interaction took place (Figure 3).

4. DISCUSSION - CONCLUSIONS

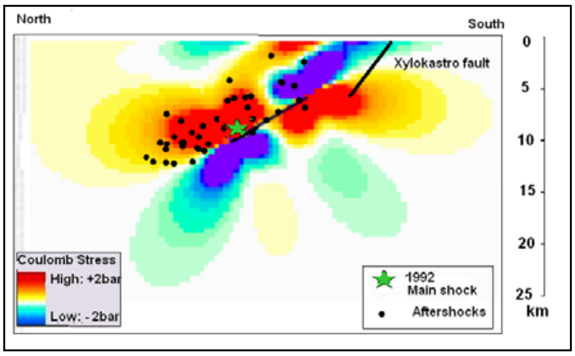
The choice of a north dipping fault for the 1992 earthquake suggested by [6, 7, 10] and adopted here is further justified by the recent high-resolution seismic reflection and multibeam bathymetric data which verify the existence of offshore north dipping faults on the northern margin of the Gulf (such as the North Eratini fault [15, 22]), although south dipping faults are in abundance [15].
In [15] it is argued that the 1995 offshore fault might be the North Eratini fault instead of the one of 30^{o} north-dip angle proposed by [4] and located 2-3km south of the Channel axis, where multibeam bathymetry and high-resolution seismic reflection data revealed no evidence of such a structure. However, a revision of the characteristics of the 1995 fault included in [7] placed this fault a little more to the west from the western edge of the N. Eratini fault and closer to the Psaromita peninsula. In contrast, considering the difficulty to clarify the geometry of active faults at significant depths, and since the N. Eratini fault appears to

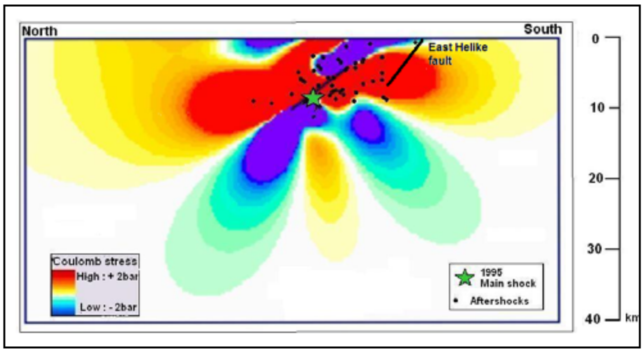
possess a shallower dip angle than the surrounding south-dipping faults (Figure 3 of [15]) it may be the one that ruptured in 1992.
With respect to the Coulomb stress change for the 1992 Galaxidi event it appears that, as in the case of the 1995 Aigion earthquake [16], the spatial aftershock pattern correlates well with regions of positive Coulomb stress changes (Figure 2a, 2b).





a. c.





b. d.

Figure 2 **a**. Map view for the 1992 event with specified faults and aftershocks locations. **b.** The 1992 cross section with the aftershock locations at depth. **c**. Map view for the 1995 event with specified faults and aftershocks locations. **d.** The 1995 cross section with the aftershock locations at depth.

With respect to the combined Coulomb stress change a rather interesting picture emerges indicating that the 1992 event probably affected the offshore fault that gave the 1995 Aigion earthquake (Figure 3). The 1992 aftershock activity outlines a fault plane broader than the one that the scaling laws predict for an earthquake of such a magnitude dipping to the north at 30^{o} [10]. However, the map projection of the aftershocks is restricted to the northern part of the positive Coulomb lobe suggesting a migration, in depth, to the NNW toward the main shock of the 1995 which is positioned to the NNE part of its respective positive Coulomb stress change. After the 1995 event took place the Coulomb stress change field due to the 1992 earthquake was partially relieved on the northern part (Figure 3). A comparison of figures 2a-d and 3 shows that the “eastern” Coulomb stress shadow of the 1995 event affected the “north” positive lobe of the 1992 event, which means that –considering the fault dip- the stress field of the 1992 was relieved in depth.

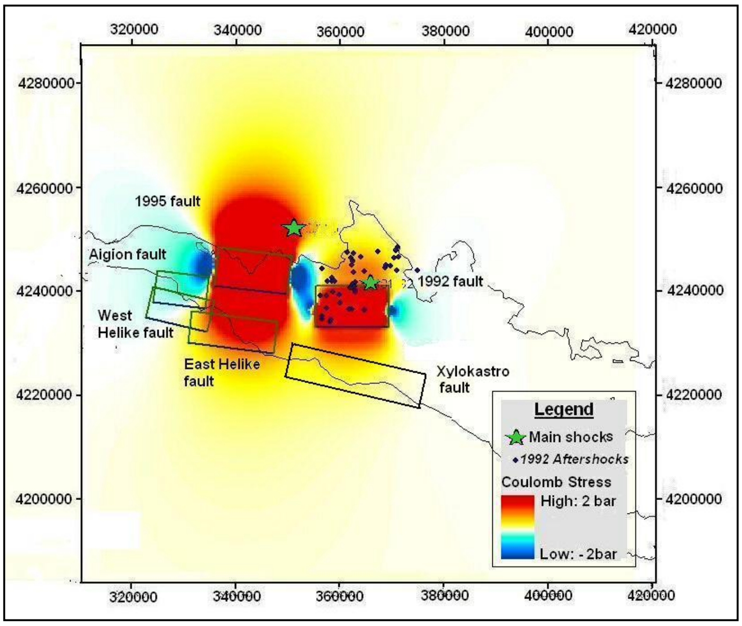
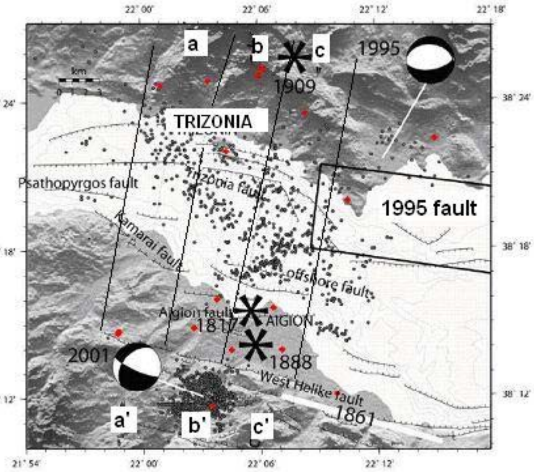


Figure 3. Combined Coulomb stress changes for the 1992 and 1995 events with aftershock activity for the 1992 event mapped.

A recent seismic swarm of December 2002, about 15km west of the Trizonia islands and at the western tip of the 1995 offshore fault was recorded and associated with a slow transient strain [5]. This may be a further indication of this westward migration of seismic activity related with the fracturing process (Figure 4a-b taken from [5]).
The active well documented faults in the region appear to be in the late part of their seismic cycle. The probability of an earthquake of M_S~6.0-6.5 is very high for the next few decades, while even the possibility of cascading events should not be entirely discarded [5]. This estimation is corroborated by the micro-seismicity in the region for the years 2000-2001 and other observations [5]. However, there is no indication that the stress change due to the 1992 event affected the stress conditions of the neighbouring active faults (Table **3**) on the south coast of the Gulf. The only exception is the Xylocastro fault which appears to be inside the positive Coulomb stress change (Figure 2a, 2b).
Regarding the Coulomb stress change of the 1995 event, the East Helike fault appears to be inside the positive stress change (~2bars) [16] (Figure 2c, 2d).
The area has significant microseismicity and such an activity has been recorded both prior to the 1992 and 1995 events. Therefore, future analysis should include in the time sequence of static stress change the major microseismicity events as well as the larger aftershocks of the two main earthquakes.
Significant (~15mm/yr) “long-term” inter-seismic displacement field for the central part of the Gulf has been well documented from more than 10 years GPS observations [1]. Recent geophysical data indicate that faults of the northern margin of the Gulf may have slip rates comparable to the south ones [15]. Geological and seismic data suggest slip rates for the Helike and Aigio faults of the order of 4-6mm/yr. If these values are doubled to account for the northern margin faults the geologically estimated deformation rates are much closer to the geodetically derived ones, at least for this part of the Gulf. Thus, it would be of interest to calculate the free surface “co-seismic” displacement field due to the estimated Coulomb stress changes and compare it to the geodetically derived one for the area of interest.



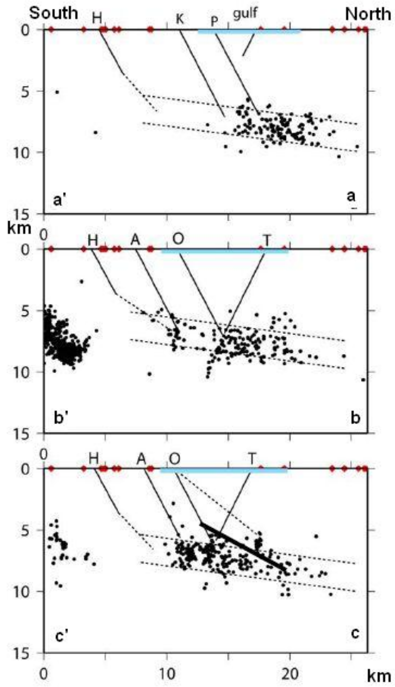


Figure 4a. Map view of the 2000-2001 seismicity. Truncated rectangle to the east corresponds to the 1995 rupture area, from [4]. Stars correspond to center of maximal intensity of historical earthquakes. Straight white segment indicates dimension of the reported surface rupture of the Heliki 1861 earthquake. Faults from [17]. Bathymetry is produced by HCMR. Focal mechanism is for the 1995 earthquake and the mainshock of the 2001 swarm. Dotted lines: cross-sections presented in **4b**. **b.** Vertical, N10°E cross-sections of the 2000-2001 seismicity on profiles aa', bd', cc' of Figure **4a**. Faults are assumed with large dip angle (60°): Aigion (A), Helike (H) faults, Kamarai (K), Psathopyrgos (P) faults, (O), offshore, and (T), Trizonia faults. A small dip offshore fault (O) is represented with a dotted line on profile cc'. The thick segment is the fault plane of the 1995 Aigion earthquake. Horizontal grey layer near the surface indicates the gulf location. The parallel, dotted lines dipping gently north outline the boundaries of the creeping, seismic layer. *Reprinted from* [5].

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