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Tectonophysics 423 (2006) 73 – 82
TECTONOPHYSICS
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The $M_W = 6.3$ 2003 Lefkada earthquake (Greece) and induced stress transfer changes

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Received 18 March 2005; received in revised form 23 September 2005; accepted 25 March 2006
Available online 1 August 2006

Abstract

A large earthquake of magnitude $M_W = 6.3$ occurred on 14 August 2003 NW of the Lefkada Island, which is situated at the Ionian Sea (western Greece). The source parameters of this event are determined using body-wave modeling. The focal depth was found equal to 9 km, the constrained focal mechanism revealed dextral strike–slip motion ($\varphi = 15°$, $\Delta = 80°$ and $\lambda = 170°$), the duration of the source time function was 8 s and the seismic moment 2.9×10^{25} dyn cm. The earthquake occurred close to the northern end of the Kefallinia transform fault, where the 1994 moderate event and its aftershock sequence were also located. The epicentral distribution of the 2003 aftershock sequence revealed the existence of two clusters. The first one is located close to the epicentral area of the mainshock, while the second southern, close to the northwestern coast of the Kefallinia Island. A gap of seismicity is observed between the two clusters. The length of the activated zone is approximately 60 km. The analysis of data revealed that the northern cluster is directly related to the mainshock, while the southern one was triggered by stress transfer caused by the main event.

Keywords: Kefallinia transform fault; NW Greece; Seismicity; Body-wave modeling; Coulomb stress

1. Introduction

The island of Lefkada is situated in the Ionian Sea (western Greece), one of the most seismically active regions in the Mediterranean (Makropoulos and Burton, 1984). The area between Lefkada and Kefallinia Islands is dominated by the Kefallinia dextral transform fault and is situated between a subduction zone to the south and a collision zone to the north. Seismological data indicate right-lateral strike–slip focal mechanisms (Anderson and Jackson, 1987; Jackson and McKenzie, 1988), in agreement with geodetic data that clearly show that the slip motion has a NNE–SSW direction (Cocard et al., 1999; Jenny et al., 2004). The seismic strain rate is in agreement with the principal horizontal axis of the total geodetic strain rate field. The Kefallinia transform fault exhibits dextral strike–slip motion at a rate of 2–3 cm $year^{-1}$ (Kahle et al., 1996; Cocard et al., 1999).

The Lefkada Island has suffered from numerous strong and destructive earthquakes since antiquity (Papazachos and Papazachou, 2002). Most events were located close to the northwestern part of the island, where severe damage was observed. Precisely, the 22 November 1704 ($M = 6.3$), the 12 October 1769 ($M = 6.7$), the 23 March 1783 ($M = 6.7$), the 28 December 1869 ($M = 6.4$) and the 27 November 1914 ($M = 6.3$)

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0040-1951/$ - see front matter © 2006 Elsevier B.V. All rights reserved.
doi:10.1016/j.tecto.2006.03.003

earthquakes were among the most important, since they caused several deaths, injuries and collapse of buildings. These disasters occurred mainly at the central and northwestern parts of the island, where fissures and landslides were also observed. This is the reason that most epicenters of the historical earthquakes are located close to the northern end of the Kefallinia transform fault, in the Ionian Sea (Kouskouna et al., 1993; Makropoulos and Kouskouna, 1994). Since these events occurred at the same epicentral area with recurrence time less than 50 years, they can be considered as repeated earthquakes. Nevertheless, large earthquakes generally occur on different segments along an active fault zone.

On the contrary, only two large events could be located close to the southwestern edge of the Lefkada Island, an area that belongs to the central part of the Kefallinia transform fault. Nevertheless, important microseismic activity is observed. The two events that are related to this area are the 22 February 1723 (M=6.7) and the 22 April 1948 (M=6.5) earthquakes. Concerning the latter event, it caused damage at the SW part of the island, while fissures and tsunami waves were observed. Two months later, on 30 June 1948, an earthquake of magnitude M=6.4 occurred at the northwestern part of the island (Fig. 1). Since the installation of the first worldwide seismological network (WWSSN), in 1964, the main events that occurred along the Kefallinia transform fault are the ones presented in Fig. 2. The two largest events that were studied using body-wave modeling are the ones that occurred in 1983 and in 2003 at the southern and the northern part of the Kefallinia transform fault, respectively.

On 14 August 2003 (05:14 GMT), a large earthquake (M_W=6.3) occurred close to the NW coast of Lefkada Island, causing some damage, landslides and ground fissures. The epicenter location of this event is 38.86 °N and 20.56 °E, in agreement with the one calculated by the National Observatory of Athens (Papadopoulos et al., in press). In the present study, the source parameters of this event are determined using body-wave modeling. The results are combined with the aftershock distribution in order to interpret the southern cluster that was activated immediately after the occurrence of the main

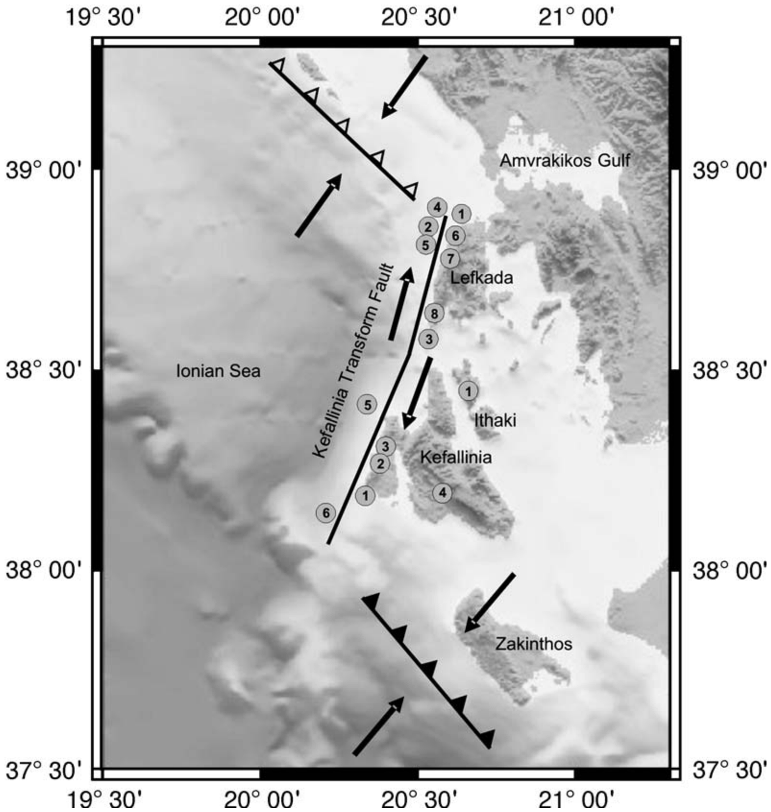

Fig. 1. Tectonic setting of the Ionian Sea. Main earthquakes since 1700. Circles close to Lefkada Island labeled 1–8 represent earthquakes that occurred on 1704, 1722, 1723, 1769, 1783, 1869, 1914 and 1948, respectively. Similarly, for Kefallinia, the event dates are 1766, 1767, 1867, 1953, 1972 and 1983 and for Ithaki 1915.

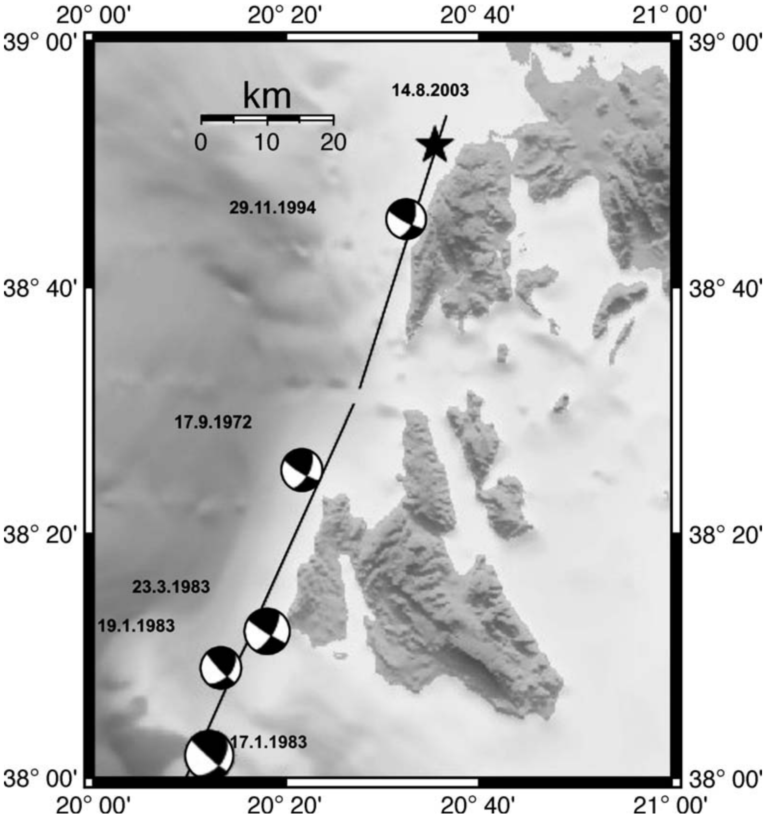

Fig. 2. Focal mechanisms of large earthquakes since 1964. The star indicates the epicenter of the 2003 Lefkada Earthquake.

event. Coulomb stress analysis is also applied in order to explain the activation of this cluster.

2. Seismicity

The study area is characterized by the occurrence of large earthquakes. The largest events that took place during the last four decades are presented in Fig. 2. The first event occurred on 17 September 1972 at the central part of the Kefallinia transform fault, close to the NW Coast of Kefallinia Island. The largest event occurred 11 years later, on 17 January 1983, at the southern part of the transform fault. This event was followed by two large aftershocks, on 19 January and 23 March. On 29 November 1994, a moderate event took place on the northern part of the transform fault, close to the west coast of Lefkada Island. The last large event is the one that occurred on 14 August 2003 at the same epicentral area. Taking into account all the above-mentioned events, it is obvious that the length of the Kefallinia transform fault is approximately 100 km, striking in an almost NNE–SSW direction. Focal mechanisms of all large events reveal right-lateral strike–slip faulting (Scordilis et al., 1985; Papadimitriou, 1988; Louvari et al., 1999). The intense seismic activity along this fault zone led to the installation of local temporal seismological networks in order to study the geodynamics of the area.

In 1989, a temporary seismological network was installed in northwestern Greece for a period of 2 months. The scope of installation of the network, which consisted of 54 portable seismological stations, was the detailed recording of microseismicity and the determination of the stress field of the area (Hatzfeld et al., 1990; Kassaras et al., 1993). During the operation of the network, 650 microearthquakes were located (Fig. 3). The seismicity is mainly concentrated close to active faults, where large earthquakes occurred in the past. Furthermore, the majority of the located events is clustered. Three clusters can clearly be distinguished along the Kefallinia transform fault. The first is located SW of Kefallinia Island, close to the epicentral area of the 17 January 1983 earthquake, while the second in the NW part of Kefallinia Island, close to the epicentral area of the 17 September 1972 earthquake. The third cluster is located close to the west coast of Lefkada Island. A part of these events is located offshore, while the others

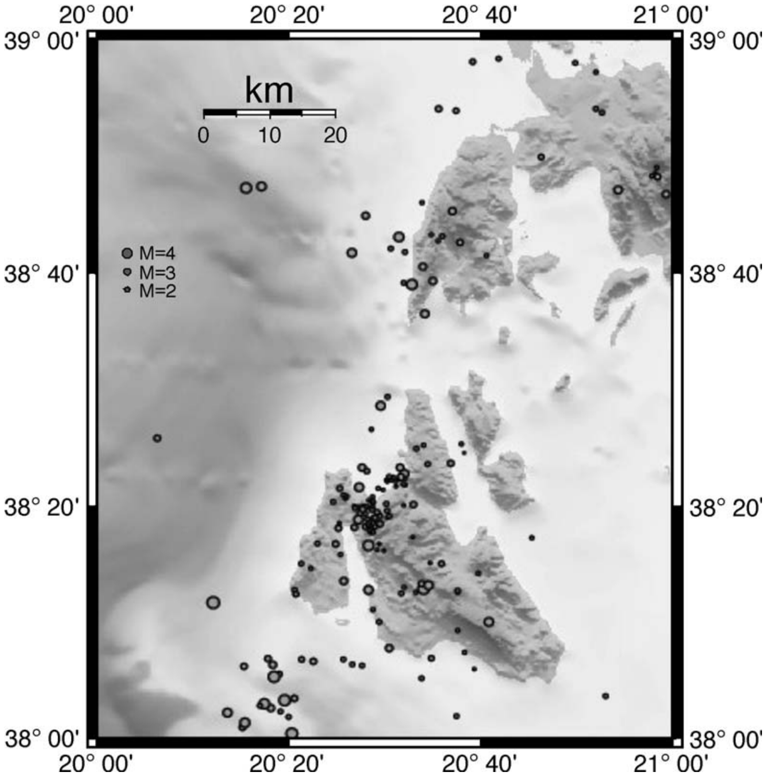

Fig. 3. Epicenters located during the 1989 experiment.

on land. The depth distribution of these events varies between surface and 20 km and reveals a sub-vertical plane dipping towards the SE (Hatzfeld et al., 1995). Fault plane solutions, which are constrained using first P-wave polarities, indicate right-lateral strike–slip faulting and P-axes trending almost NE.

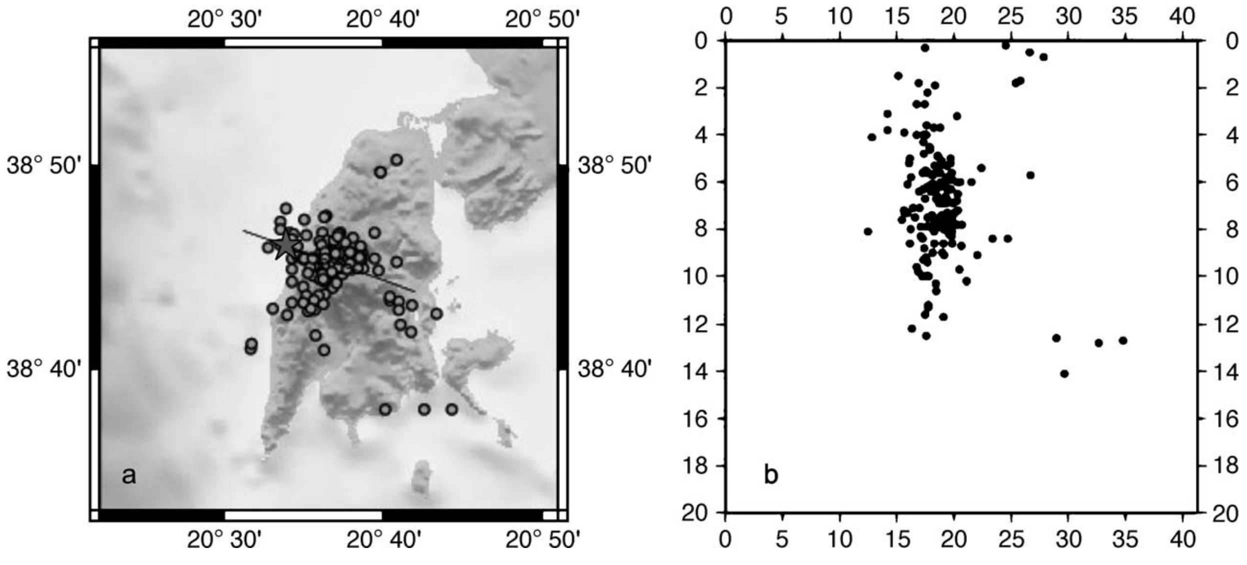

Fig. 4. (a) Aftershocks of the 1994 earthquake sequence recorded by the digital seismological network installed by the University of Athens. The star indicates the epicenter of the 29 November 1994 earthquake. The line indicates the direction of the cross-section presented in (b). (b) Seismic cross-section along N120° direction, as indicated in (a). An almost vertical depth distribution is observed.

On November 29 1994 (14:30 GMT), a moderate earthquake of $M_W = 5.1$ took place close to the western coast of Lefkada Island, followed by a large aftershock on December 1 (07:17 GMT, $M = 4.8$). The aftershock sequence (Fig. 4a) was recorded by a temporary seismological network installed by the University of Athens. This was the first digital network installed in the area and consisted of six seismographs equipped with three component seismometers. All stations were installed on the island for a period of 3 weeks, following the occurrence of the mainshock, in order to record the aftershock sequence. The majority of the located events is concentrated in the west-central part of the island. A cross-section (Fig. 4b), performed in E–W direction, indicates that focal depths vary between surface and 12 km with an almost vertical distribution (Makropoulos et al., 1996). The constrained focal mechanisms revealed dextral strike–slip motion on a NNE–SSW east-dipping nodal plane, consistent with the direction of the observed co-seismic surface fissures.

Finally, during the summer of 1995 a seismological network was installed east of Lefkada Island, on the continental part of western Greece. Intense seismic activity was observed around Amvrakikos Gulf, while very low seismicity was recorded along the northern part of the Kefallinia transform fault (Haslinger et al., 1999).

3. Source parameters of the 14 August 2003 Lefkada earthquake

The source parameters of the 2003 Lefkada earthquake are determined using body-wave modeling. The calculation of a synthetic seismogram $W(t)$ is based on the convolution of four linear operators:

$$W(t) = U(t)^{*}S(t)^{*}I(t)^{*}Q(t)$$

where U is the far-field displacement, S is the source time function, I is the instrument response and Q is the attenuation. The far-field wave displacement, recorded at azimuth φ and epicentral distance Δ, is calculated using the formula (Kanamori and Stewart, 1976; Aki and Richards, 1980):

$$U(t) = \frac{M_o}{4\pi\rho c^3}\frac{g\Delta}{a}C(i_o)R(\phi, i_h)$$

where U is the vertical, radial or transverse displacement in case of P-, SV- or SH-waves, respectively, ρ is the density, c the velocity for P- or S-waves, φ is the azimuth, α the Earth radius, i_h is the angle of incidence at source and i_o is the angle of incidence at the receiver, $g(\Delta)$ is the geometrical spreading, C is the free-surface effect, and R is the radiation pattern. In order to calculate synthetic seismograms, recordings of stations at distances between 30° and 90° are selected, to avoid complexities due to the upper mantle response and to core phase effects. The instrument response $I(t)$ is removed from the selected recordings and a Butterworth band-pass filter is applied. For the estimation of the attenuation factor $Q(t)$, t^* is considered 1 and 4 for P- and S-waves, respectively (Futterman, 1962).

Initially, synthetic waves are calculated by trial-and-error method to determine the focal mechanism, the focal depth, and the seismic moment for a single trapezoidal source time function. The next step is to apply inversion with a fixed fault mechanism in order to estimate the complexity of the source that represents the sub-event distribution over the fault plane (Bezzeghoud et al., 1986). The applied technique is similar to the one described by Kikuchi and Kanamori (1982, 1991). The complicated waveforms recorded by different stations imply a complex rupture process, a fact that was revealed by the inversion that was performed. The duration of the elementary pulses is longer in the stations that are situated north of the epicenter of the mainshock, comparing to those situated to the south. These facts, combined with the hypocenter location, which lies close to the northern end of the rupture fault, imply a probable directivity towards the south. The final calculation of the source parameters was performed using the source time function that was determined by the recordings of the ARU station. The result revealed a double source with duration equal to 8 s. The application of body-wave modeling for the 2003 Lefkada earthquake (Fig. 5) revealed right-lateral strike–slip focal mechanism ($\varphi = 15°$, $\Delta = 80°$ and $\lambda = 170°$). The focal depth is equal to 9 km and the seismic moment 2.9×10^{25} dyn cm. Taking the velocity rupture equal to 3 km/s, the fault length is estimated 24 km and the coseismic slip 40 cm.

The aftershock sequence of the 2003 Lefkada earthquake is presented in Fig. 6. Epicenters were located by the permanent seismological network of Patras University (http://seismo.geology.upatras.gr). More than 600 aftershocks were located until 30 September 2003. Two clusters, activated almost simultaneously after the occurrence of the mainshock, can clearly be distinguished. The first one lies close to the epicentral area of the mainshock, while the second is located to the south, close to the northwestern coast of the Kefallinia Island. A ‘gap’ is observed between the two clusters. The length of the area covered by the first cluster is approximately 24 km, while the hypocentral distribution presents an almost vertical direction.

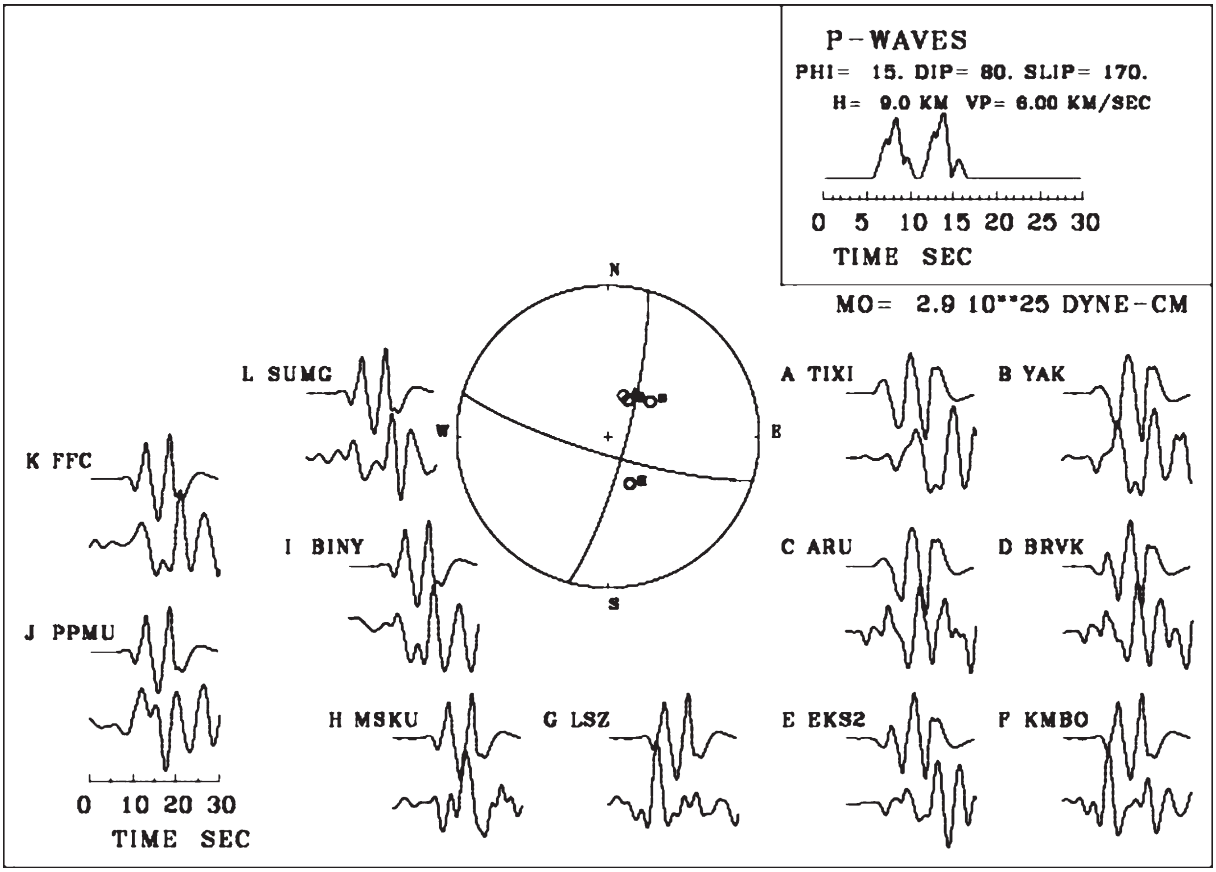

Fig. 5. Body-wave modeling of P teleseismic waves.

Similar results for the aftershock sequence were obtained by a local temporary seismological network installed by the University of Thessaloniki (Karakostas et al., 2004). The total length of the activated area was estimated greater than 60 km. The epicentral distribution of the 1994 earthquake and the central (main) part of the northern cluster of the 2003 aftershocks cover the same epicentral area. Since 1995 the seismicity in the area progressively decreased, leading to a clear seismic quiescence 1 year before the occurrence of the 2003 mainshock. Seismic quiescence is due to a quasi-static decrease in stress caused by stress relaxation in the volume around the mainshock area and is considered as a seismic precursor phenomenon (Wyss and Habermann, 1988). Thus, after the 1994 earthquake sequence, a period of seismic quiescence was initiated, which can be considered as an earthquake preparation phase of the 2003 Lefkada earthquake (Papadimitriou, 2004).

Taking into account the focal mechanism of the mainshock and the general NNE–SSW direction of the aftershock distribution, the most likely activated fault plane is the one with right-lateral strike–slip motion, vertical direction and weak SE dip. This direction is also in agreement with the bathymetry of the area and the steep coastline of the western part of the Lefkada Island.

4. Coulomb stress changes

Local triggering effects that can clearly be observed are studied using classical concepts of stress transfer. The Coulomb failure criterion satisfies conditions similar to those of friction on a pre-existing surface, where failure occurs on a plane when the stress exceeds a specific value. Coulomb failure stress changes caused by a large earthquake explain satisfactorily the spatial distribution of the aftershocks, as well as the existence of distant events triggered by stress changes of less than 1 bar, even of the order of 0.1 bar, that can cause microseismicity rate changes (Reasenberg and Simpson, 1992; Toda et al., in press). In cases of larger stress changes another main event may be generated. A well-known example is the Landers earthquake, where approximately 3 h after its occurrence, the Big Bear earthquake was triggered (Stein et al., 1992; King et al.,

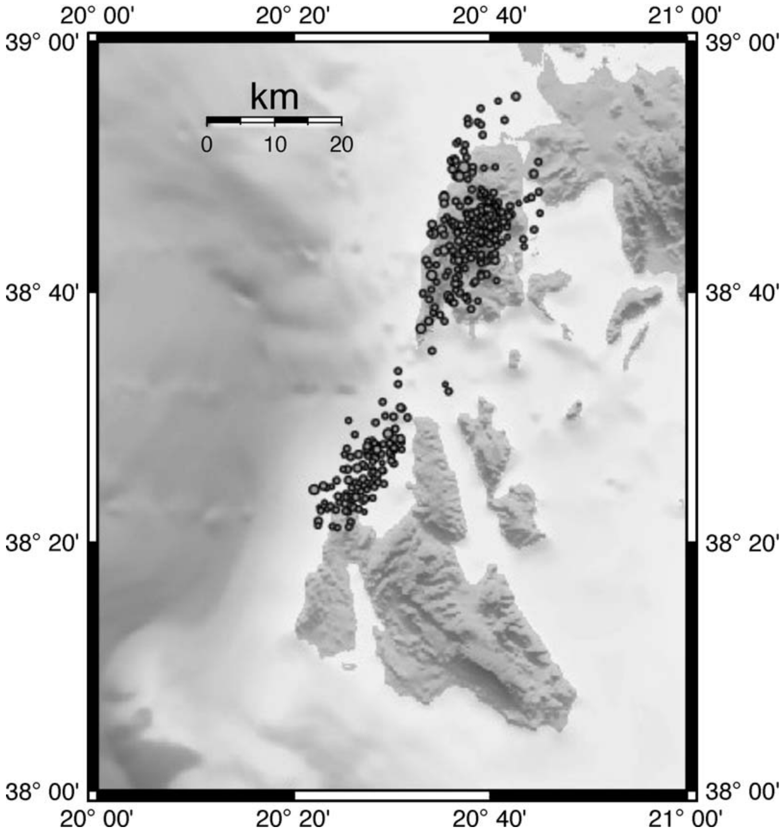

Fig. 6. Selected aftershocks of the 2003 Lefkada earthquake located by the permanent seismological network of the Patras University.

1994). These observations led us to apply this technique in the case of the Lefkada 2003 earthquake in order to examine whether the existence of the second cluster, located approximately 20 km south of the rupture zone, can be attributed to stress transfer changes due to the mainshock.

Earthquake-induced stress changes, applied to a fault plane, are estimated using dislocation calculations that simulate static earthquake slip in an elastic half-space (Okada, 1992). This technique is performed using the Coulomb Failure Stress, ΔCFS (Harris, 1998; Harris and Simpson, 2002)

$$\Delta\mathrm{CFS} = \Delta\tau + \mu(\Delta\sigma + \Delta p)$$

where $\Delta\tau$ is the change in shear stress on the failure plane, μ is the coefficient of friction, $\Delta\sigma$ is the change of normal stress (positive for tension) and Δp is the pore pressure change. The advantage of using changes in stress is that most often the absolute values of the stress are not known, but stress change values can be calculated using the geometry and the slip direction of an earthquake rupture. For this calculation, μ is considered constant and is assumed not to change as a function of slip on the rupture plane. The Skempton's coefficient B_k that varies between 0 and 1 is used to incorporate pore fluid effects. The effective coefficient of friction,

$$\mu' = \mu(1 - B_k)$$

is used to calculate the Coulomb failure criterion (Rice, 1992). The following equation can be obtained with the assumption that the medium is homogeneous and isotropic (Harris, 1998):

$$\Delta\mathrm{CFS} = \Delta\tau + \mu'\Delta\sigma$$

The locations with positive values of ΔCFS are considered to be advanced toward failure and the locations with negative values are delayed from failure and are commonly called stress shadows.

Coulomb stress changes are calculated for the case of the Lefkada 2003 earthquake (Fig. 7), using an effective coefficient of friction μ' equal to 0.5 and the source parameters calculated in the present study. The calculation was performed considering that the length of the

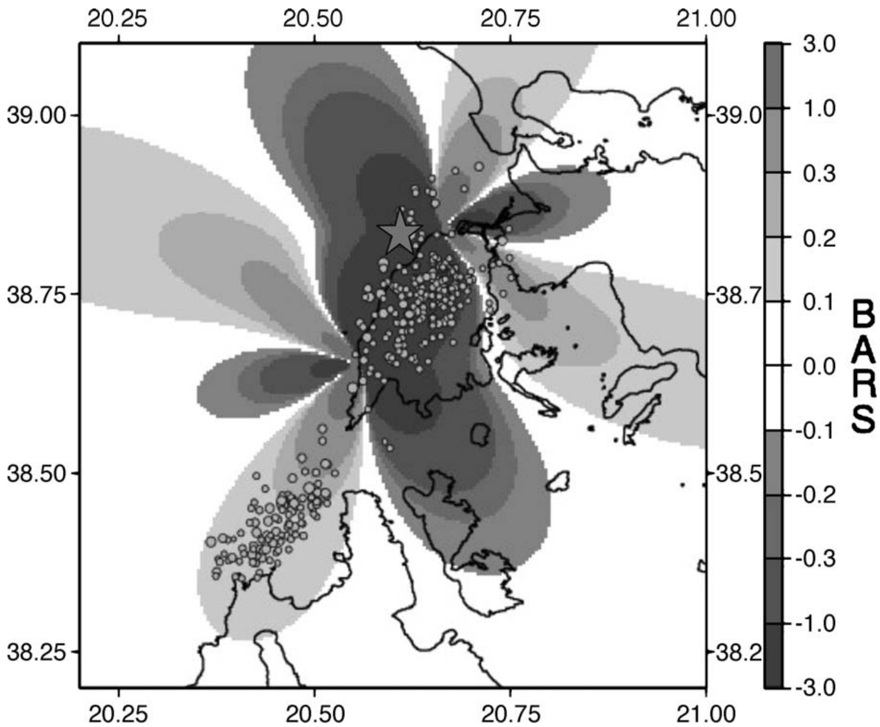

Fig. 7. Static Coulomb stress changes caused by the 2003 Lefkada earthquake. The star represents the epicenter of this mainshock. Aftershocks are also presented.

main rupture fault is 24 km and its width extends from 2 to 12 km and the associated slip is 40 cm. In the same figure, the aftershock locations are also plotted. The first cluster that is located close to the main rupture fault is successfully predicted on the basis of the Coulomb failure criterion. More specifically, the events occur where Coulomb stresses have risen. The stress pattern is controlled by both the initiation and termination ends of the main rupture. The activity along the rupture fault is successfully predicted, as well as the off-fault activity by the northern and southern lobes. Furthermore, the spatial distribution is not only restricted along the western coast of the Lefkada Island, but it is also expanded in the northeastern part of the island. This observation is successfully predicted by the northwestern lobe.

As already mentioned, a second cluster is located about 20 km south of the main rupture, while a “gap” of seismic activity is observed between the two clusters. The southern cluster is located within an area of high seismicity, where the 1972 moderate earthquake also occurred. Furthermore, the second cluster of the 1989 experiment was located close to this area. This cluster consists of events that are characterized by low magnitudes, while their number during the first hours after the occurrence of the mainshock is significantly smaller, compared to the northern cluster. Furthermore, all large aftershocks for the same time period were located at the area covered by the northern cluster. Additionally, no damage was observed at the north-western part of the Kefallinia Island. All events of the southern cluster occurred within the southern lobe that was calculated by the Coulomb stress changes. This area is loaded with additional stress, larger than 0.15 bar. Thus, it is reasonable to propose that the small perturbation of the stress field caused by the 2003 Lefkada earthquake played a major role in triggering the activation of this cluster by producing microseismicity rate changes. An additional fact is the occurrence of the 16 November 2003 moderate earthquake ($M_W = 5.1$), located south of this cluster, during a time period when the northern cluster presented low seismic activity.

An alternative explanation for the existence of this cluster was proposed by Zahradnik et al. (in press). According to these authors, the Lefkada earthquake sequence was interpreted as a double event, implying that the southern cluster is the aftershock sequence of an independent main event. The results of body wave modeling, performed in our study, did not reveal the existence of two clear independent main events. The complexity of the recorded teleseismic waveforms is probably due to the directivity of the seismic source rupture process towards the south.

5. Conclusions

The 2003 Lefkada earthquake occurred close to the northern end of the Kefallinia transform fault that is characterised by dextral strike–slip motion. The source parameters of this event were calculated by applying body wave modelling using teleseismic recordings. The analysis provided a right-lateral strike–slip focal mechanism (strike = 15°, dip = 80° and rake = 170°) at a centroid depth of 9 km, consistent with the NEIC and Harvard determinations. The obtained source presents complexities and consists of two elementary sources, and its total duration is 8 s, implying that the length of the activated fault is 24 km, in agreement with the length of the activated aftershock area. The hypocenter of the mainshock is located to the deep northern edge of the activated fault, with a probable directivity towards the south.

The aftershock activity is characterised by the existence of two clusters. The first is directly related to the main ruptured area and is aligned in an almost NNE–SSW direction. The length of this cluster is approximately 24 km, while its depth distribution is almost vertical. The aftershock spatial distribution presents an expansion in the northeastern part of the island which was well predicted by Coulomb stress changes.

The second cluster is located south of the rupture zone, close to the northwestern coast of the Kefallinia Island. A seismic "gap" is observed in the area between the two clusters. Clusters have also been observed elsewhere in Greece (Burton, 1993; Burton et al., 1995). The 1980 Volos earthquake sequence (Central Greece) can be mentioned, where two clusters were observed and interpreted in terms of potential seismic hazard. Burton et al. (1991) examined whether the entire block structure between the two clusters should rupture partially or as a whole. Data analysis revealed that the fault area was partially activated, mainly close to its two edges, with a "gap" of about 8 km between them. The 30 April 1985 Volos earthquake occurred within this area (Papadimitriou et al., 1993).

Coulomb stress analysis was performed, using the source parameters calculated in the framework of the present study, and revealed that the southern cluster is triggered by the main event, as an effect of stress transfer loading. The area was loaded with additional stress greater than 0.15 bar. These observations suggest that this area of enhanced Coulomb stress, where lack of seismicity is also observed, should be regarded as candidate for a future main event.

Acknowledgments

The Coulomb stress changes analysis was performed using the DLC Software provided by R.W. Simpson (USGS). Figures were plotted using the GMT software (Wessel and Smith, 1995). The present study was funded through the program EPEAEK II in the framework of the project "Pythagoras—Support of University Research Groups" with 75% from European Social Funds and 25% from National Funds.

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