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Current accelerating seismic excitation along the northern boundary of the Aegean microplate

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Abstract

According to previous observations [Geophys. Res. Lett. 27 (2000) 3957], the generation of large ($M \geq 7.0$) earthquakes in the western part of the north Anatolian fault system (Marmara Sea) is followed by strong earthquakes along the Northern Boundary of the Aegean microplate (NAB: northwestermost Anatolia–northern Aegean–central Greece–Ionian islands). Therefore, it can be hypothesized that a seismic excitation along this boundary should be expected after the occurrence of the Izmit 1999 earthquake ($M=7.6$). We have applied the method of accelerating seismic crustal deformation, which is based on concepts of critical point dynamics in an attempt to locate more precisely those regions along the NAB where seismic excitation is more likely to occur. For this reason, a detailed parametric grid search of the broader NAB area was performed for the identification of accelerating energy release behavior.

Three such elliptical critical regions have been identified with centers along this boundary. The first region, (A), is centered in the eastern part of this boundary (40.2°N, 27.2°E: southwest of Marmara), the second region, (B), has a center in the middle part of the boundary (38.8°N, 23.4°E: East Central Greece) and the third region, (C), in the westernmost part of the boundary (38.2°N, 20.9°E: Ionian Islands). The study of the time variation of the cumulative Benioff strain in two of the three identified regions (A and B) revealed that intense accelerating seismicity is observed especially after the occurrence of the 1999 Izmit mainshock. Therefore, it can be suggested that the seismic excitation, at least in these two regions, has been triggered by the Izmit mainshock.

Estimations of the magnitudes and origin times of the expected mainshocks in these three critical regions have also been performed, assuming that the accelerating seismicity in these regions will lead to a critical point, that is, to the generation of mainshocks.

Keywords: Accelerating seismic deformation; Aegean; Earthquake prediction

1. Introduction

The Aegean area (34°N–43°N, 19°E–30°E) is seismically the most active area of western Eurasia. This is mainly due to the existence and fast southwestward motion of the Aegean microplate with respect to the Eurasian plate (McKenzie, 1970, 1978; Oral et al., 1995; Papazachos, 1999). In the south, this microplate is bounded by the Hellenic Arc (Zante–Crete–Rhodos) where it overrides the east Mediterranean lithosphere (front part of the

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African lithospheric plate) and where shallow and intermediate depth earthquakes, with magnitudes up to about $M = 8.0$, occur (Fig. 1). The northern boundary of the microplate has an ENE–WSW trend (West Marmara–Northern Aegean–Central Greece–Ionian Islands) where shallow earthquakes, with magnitudes up to $M = 7.5$, occur with strike–slip dextral or normal faulting. It consists of the western part of the Northern Anatolian Fault Zone (NAF in Fig. 1) and its westward continuation, that is, the Northern Aegean Trough (NAT in Fig. 1). Although there is no clear continuation of the strike–slip motion from the Northern Aegean Trough to the Ionian Sea, there is a recognizable band of strong earthquake epicenters, surface breakages of normal faults and rupture zones (Papazachos et al., 1993, 1999; Goldsworthy et al., 2002). These observations and the implications of the model proposed by Papazachos (1999), who used all available seismological and GPS evidence for the Aegean–Anatolia interaction, suggest that this band, which becomes indistinct in the flysch belts of the Pindos mountains, may probably connect the strike–slip faulting in the offshore NAT to the east with the strike–slip faulting in the Cephalonia Transform Fault to the west (CFT in Fig. 1) and its northern edge in the Amvrakikos

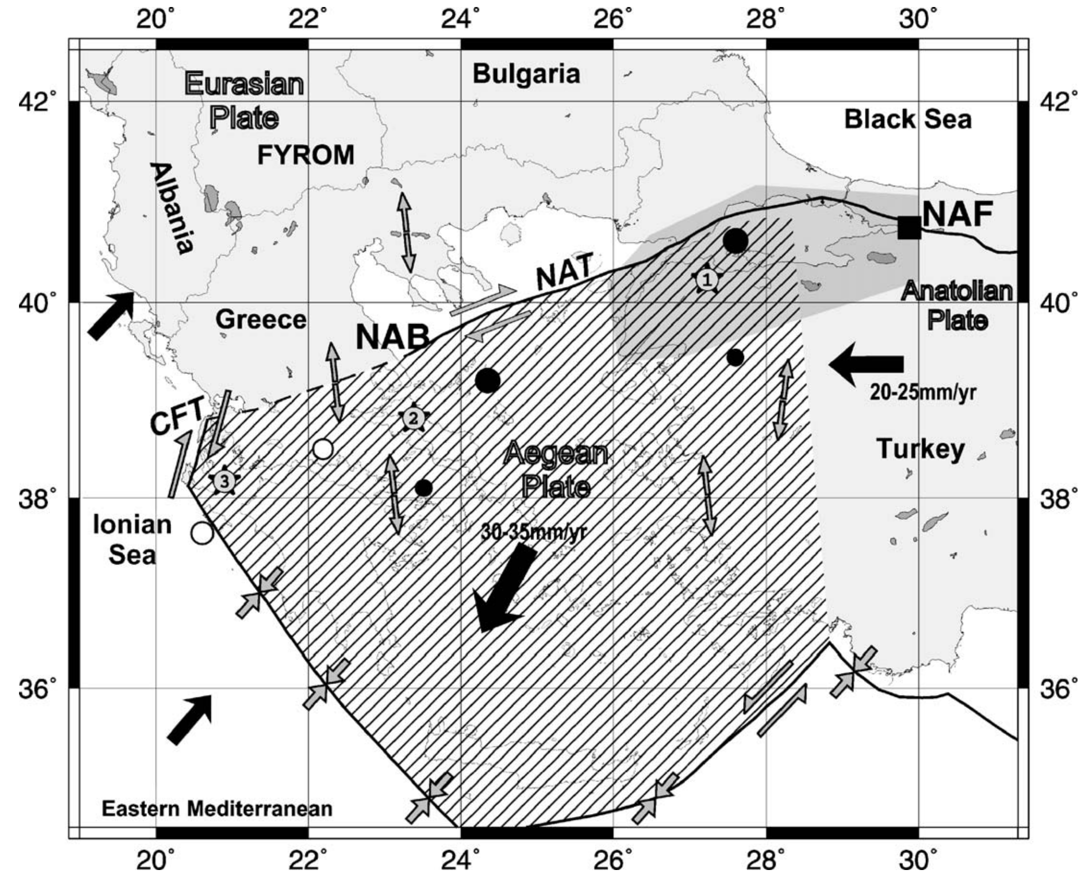

Fig. 1. Main tectonic features of the broader Aegean area. The black square denotes the epicenter of the Izmit 1999 large earthquake. Stars are the centers of the three critical regions along the northern boundary of the Aegean microplate and the number inside each center corresponds to the code number of Table 1. Large circles show epicenters of the largest shock and small circles show epicenters of the second largest shock, which occurred up to now in each sequence. Black and white circles show epicenters of shocks, which occurred after or before the Izmit 1999 earthquake, respectively. Local stresses and fault motions are shown with grey arrows, while average plate motions with respect to the stable Eurasia are denoted with solid arrows.

area, which is considered as a triple-junction (King et al., 1993). The eastern boundary of the plate (western Turkey) is an extensional area which marks the change that occurs from the Anatolia rotation to the Aegean translation and includes the central and southern part of the western coasts of Turkey and the neighbouring Greek islands (Papazachos, 1999). In this area earthquakes, with magnitudes up to $M = 7.5$, occur.

It has been observed (Papazachos et al., 2000) that large earthquakes ($M \geq 7.0$) which occurred during the last four centuries in the western part of the north Anatolian fault system (Marmara Sea) triggered strong seismicity in the Aegean within a time period of about 5 years and that this triggered seismicity started by the generation of strong earthquakes along the northern boundary of the Aegean plate. For this reason, and since the Izmit earthquake of August 17, 1999 ($M = 7.6$, 40.8°N, 30.0°E) is such a large earthquake which occurred in the western part of the North Anatolian Fault (NAF), excitation of strong seismic activity should be expected along NAB. Two strong earthquakes have already occurred in the Aegean after the Izmit earthquake, a destructive one in central Greece (7.9.1999, $M = 5.9$) and a damaging one in north Aegean (26.7.2001, $M = 6.4$) but the first of these events is relatively small in comparison with such earthquakes in all 12 previous observed cases (Papazachos et al., 2000). For example, one of the last previous large earthquakes in NAF (18.3. 1953, $M = 7.4$, 40.0°N, 27.5°E) was followed by a large earthquake in the westernmost part of the NAB (12.8.1953, $M = 7.2$, 38.1°N, 20.6°E) and by another large earthquake in the central part of NAB (30.4.1954, $M = 7.0$, 39.3°N, 22.3°E). Hence, it is possible that such strong earthquakes will occur in the near future at some places of this boundary.

Therefore, it is of interest to study the northern boundary of the Aegean microplate (NAB) for identification of regions that are in seismic excitation and may lead to the generation of strong mainshocks. A proper method for this is the "accelerating seismic crustal deformation method" (Papazachos and Papazachos, 2000, 2001), which is based on concepts of the critical point dynamics (Sornette and Sammis, 1995; Jaume and Sykes, 1999).

2. Method and data

The method applied to identify foci of strong earthquakes along the northern boundary of the Aegean microplate (NAB) relies on an approach where regions of accelerating seismic crustal deformation (critical regions) are identified (Papazachos and Papazachos, 2001). This method is based:

(a) On the following equation which expresses the power–law that relates the cumulative Benioff strain, $S(t)$, released by the intermediate magnitude shocks (preshocks), with the origin time, t_c, of the mainshock:

$$S(t) = A + B(t - t_c)^m \quad (1)$$

where t is the time to the mainshock and A, B, m are model parameters (Bufe and Varnes, 1993). Usually B is negative, m is positive and smaller than 1 (for acceleration) and A is the total Benioff strain including the strain of the mainshock (Varnes, 1989). The use of cumulative Benioff strain, which is the square root of seismic energy and can be easily estimated from the moment magnitude of the shocks, leads to more stable solutions of Eq. (1) and predictions of the time of the oncoming mainshock than the use of the cumulative seismic moment release (Bufe and Varnes, 1993; Jaume and Sykes, 1999). The term "intermediate magnitude shocks" refers to those shocks with magnitudes ~ 2 magnitude units smaller than the mainshock magnitude (Bowman et al., 1998), whereas the choice of the appropriate cutoff preshock magnitude for the identification of a critical region seems to depend on the magnitude of the oncoming mainshock (Papazachos, in press).

(b) On minimizing a curvature parameter, C, which is defined as the ratio of the root mean square error of the power-law fit (relation (1)) to the corresponding linear fit error (Bowman et al., 1998) and practically quantifies the degree of deviation of the released Benioff strain from linearity (acceleration); small C values (≤ 0.7, Bowman et al., 1998) ensure that the accelerating seismicity law (1) describes the data much more adequately than the standard linear time variation, in order to avoid misidentifications.

(c) On additional constraints expressed by several relations between the parameters involved (Papazachos and Papazachos, 2000, 2001). These relations, which have recently been redefined by using global data (Papazachos et al., submitted for publication(a)), relate: (a) the radius, R (in km), of the circle with area equal to the area of the elliptical region with the mainshock magnitude, M, and the mean deformation rate in the area, s_r (in $J^{1/2}$/year and per 10^4 km^2), (b) the duration of the preshock sequence, t_p (in years), with s_r, (c) the mean rate of deformation during the accelerating deformation period, A/t_p, with the long-term rate of seismic deformation, S_r, and (d) the mainshock magnitude, M, to the average magnitude of the three largest preshocks, M_{13} (relations (2)–(5)).

$$\log R = 0.42M - 0.34\log s_r + 1.52 \quad \sigma = 0.11 \qquad (2)$$

$$\log t_p = 3.60 - 0.40\log s_r \quad \sigma = 0.15 \qquad (3)$$

$$\log(A/t_p) = 1.01\log S_r \quad \sigma = 0.02 \qquad (4)$$

$$M = M_{13} + 0.60 \quad \sigma = 0.15. \qquad (5)$$

In order to quantify the compatibility of these relations with observations, a parameter P was defined (Papazachos and Papazachos, 2001), which is the average value of the probabilities that each of these four parameters (R, t_p, A, M) attains a value close to its expected one, using a Gaussian probability density function based on the deviations reported in Eqs. (2)–(5). Furthermore, Papazachos et al. (2002a) have defined a quality measure, q, given by the relation:

$$q = \frac{P}{mC} \qquad (6)$$

in an attempt to simultaneously evaluate: (a) the compatibility of an accelerating seismic deformation with the behavior of past real preshock sequences (large P), (b) the deviation of the variation with time of the seismic deformation from linearity (small C) and (c) the degree of seismic acceleration (small m). Investigation of preshock sequences of strong mainshocks ($M \geq 6.4$), which occurred in the Aegean since 1950, has led to the adoption of the following cut-off values:

$$C \leq 0.60, \; m \leq 0.35, \; P \geq 0.45, \; q \geq 3.0 \qquad (7)$$

The obtained results for the Aegean show a typical average value of $m = 0.29$ (which was also used in the present study), in accordance with theoretical considerations (Ben-Zion et al., 1999; Rundle et al., 2000). The typical values for the other parameters are $[C] = 0.44$, $[P] = 0.53$ and $[q] = 4.9$.

Application of this method in many preshock sequences in Greece and other areas (Papazachos et al., submitted for publication(a)) has shown that the seismic excitation in an elliptical critical region centered at a certain point depends on q and on other measurable quantities such as the magnitude M of the expected mainshock, the number, N, of valid solutions (solutions which fulfill relation (7)) with different elliptical regions centered at the same point and the percentage, λ, of the neighboring points of the grid for which valid solutions exist. Thus, a quantity, K, called excitation indicator, is defined by the relation:

$$K = q\frac{M}{M_c}\sqrt{\frac{N}{N_c}}\lambda \qquad (8)$$

where M_c is a constant magnitude (e.g. $M = 7.0$) and N_c is a constant number (e.g. $N_c = 10$). K attempts to modify q in order to further focus on larger mainshocks (large M values) and regions with a large number of solutions (high N values) and a high number of solutions in neighboring regions (high λ values). This is in accordance with the results of Yang et al. (2001) who, studying critical regions in New Zealand and China, found that mainshocks may occur in the vicinity of the area with the higher density of obtained solutions of accelerated deformation.

To apply this method for identification of an elliptical preshock (critical) region an algorithm has been developed (Papazachos, 2001). According to this algorithm, earthquakes (preshocks) with epicenters in an elliptical region centered at a certain geographical point (assumed epicenter of the mainshock) are considered and the parameters of relation (1) as well as the curvature parameter, C, are calculated. Calculations are repeated by applying a systematic search of a large set of values for the azimuth, z, of the large ellipse axis, its length, a, ellipticity, e, and the time, t_s, since when accelerating seismic deformation started, for the magnitude, M, and the origin time, t_c, of the mainshock. These computations are repeated for a grid of points in which the investigated area is separated with the desired density (e.g. 0.2°NS, 0.2°EW). The geographical point which corresponds to the best solution (largest value of K) is considered as a first approximation of the epicenter of the expected mainshock and the magnitude, which corresponds to this solution, is considered as the magnitude of the mainshock. The finally adopted epicenter is based on additional information, such as the location of seismic faults in the region. The origin time of the oncoming mainshock is calculated by direct application of relation (1) with constant value of m (= 0.29) and by an additional approach proposed by Papazachos et al. (2001), which is based on an observable preshock excitation and permits calculation of t_c. The errors of the estimations (predictions) are less than 120 km for the epicenter, ± 0.5 for the magnitude and ± 2 years for the origin time of the ensuing mainshock (Papazachos et al., 2002b).

The previously described procedure has been applied to identify elliptical regions that are currently in a state of accelerating seismicity along the northern Aegean boundary by using data (epicenter, origin times and magnitudes of shocks) for earthquakes which occurred between 1965 and 30 September 2002 and are complete for $M \geq 4.5$. The data have been taken from the catalogue of Papazachos et al. (2003b). All magnitudes are either original or equivalent moment magnitudes derived from local magnitudes using appropriate transformations (Papazachos et al., 1997; Margaris and Papazachos, 1999; Papazachos et al., 2002c) and their standard errors are of the order of 0.3. The equation $\log E = 1.5M + 4.7$ has been used to calculate the seismic energy, E, from the moment magnitude (Vassiliou and Kanamori, 1982; Kanamori et al., 1993), which in turn is used to calculate the Benioff strain, $S(\sqrt{\sum_i E_i})$. It should be noted that Gross and Rundle (1998) have found that the time-to-failure models are robust enough to apply to catalogues with magnitude uncertainties of several tenths of a unit.

3. Results

The previously described method has been applied to identify elliptical critical regions along the northern boundary of the Aegean plate. Three such regions have been identified and their centers are indicated by stars in Fig. 1. The first one is located in the northeastern part of the boundary (westernmost part of the Anatolian Fault), the second in the central part of the boundary (eastern central Greece) and the third in the western part of the boundary (Ionian Sea). The two largest, already occurred, shocks (preshocks) in

Table 1
Information on the three elliptical regions, which are located at the northern boundary of the Aegean microplate

No.	Region	φ, λ	M	t_c	a	z	e	M_{min}	n	t_s	q	K
1	SW Marmara	40.2, 27.2	6.4	2004.1	161	0	0.90	4.7	29	1984	7.5	8.4
		40.7, 27.6	5.9	20.9.1999								
		39.4, 27.7	5.3	22.6.2001								
2	E Central Greece	38.8, 23.4	6.8	2004.8	236	0	0.90	5.1	21	1984	9.1	7.2
		38.1, 23.5	5.9	7.9.1999								
		39.1, 24.4	6.4	26.7.2001								
3	Ionian Islands	38.2, 20.9	7.0	2005.2	312	90	0.95	5.3	28	1984	3.1	0.6
		38.4, 22.2	6.4	15.6.1995								
		37.6, 20.6	6.6	18.11.1997								

The code number, N_0, the name and the geographic coordinates of the center of each region are shown in the first three columns. M and t_c are the magnitude and the origin time of the expected main shock. a (in km), z and e are the length of the large axis, the azimuth of this axis and the ellipticity of the region. M_{min} is the minimum magnitude considered and n is the number of shocks. t_s is the starting year of the accelerated deformation sequence, q is the quality measure and K is the excitation indicator (see text for explanation). In the second and third line of each group, the epicenter coordinates, the magnitude and the date of the two largest preshocks of each sequence are presented.

each of these three currently active elliptical regions are denoted by small circles.

Table 1 gives information for the three critical regions and for the epicenter coordinates, magnitudes and dates of the two largest shocks of each preseismic sequence that have already occurred. The name of the region, the geographic coordinates (φ, λ) of the center of the elliptical region, the magnitude, M, and the origin time, t_c, of the expected mainshocks are given in the first five columns of this table. In the remaining columns of this table, information is given on the parameters of the corresponding critical region, that is, on the length, a (in km) and azimuth, z, of the large axis of the elliptical region, the ellipticity, e, of the region, the minimum preshock magnitude, M_{min}, the number of preshocks, n, the start year, t_s, of the seismic sequence, the quality factor, q, and the excitation indicator, K. It is observed that the magnitudes of the three expected mainshocks are 6.4, 6.8 and 7.0 and the three corresponding expected occurrence times, are 2004.1, 2004.8 and 2005.2 years for the eastern (A), central (B) and western (C) critical region, respectively. It is of interest to note that region A has already been identified in a previous work (Papazachos et al., 2002d) and the results with respect to the center of the elliptical region and the magnitude of the expected mainshock are almost identical; the epicenter coordinates of the expected mainshock had been found to be 40.0°N–27.4°E, whereas its magnitude was estimated to 6.4. However, the estimated origin time of this main shock differs (2002.5 in Papazachos et al., 2002d and 2004.1 in the present work). This discrepancy is obviously due to the fact that the estimation of the origin time of the expected mainshock is very sensitive to the behavior of accelerating deformation during the last phase of this physical process. Moreover, it suggests that the real uncertainties in the estimation of the origin time are perhaps higher than previously estimated expected errors ($\pm$2.0 years, Papazachos et al., 2001).

Fig. 2a,b and c shows the time variation of the cumulative Benioff strain for the three sequences. Large arrows indicate the Izmit 1999 mainshock origin time in all three plots, whereas in Fig. 2b two small arrows mark the occurrence times of the two mainshocks mentioned in the Introduction (7.9.1999 M=5.9; 26.7.2001 M=6.4). The accelera-

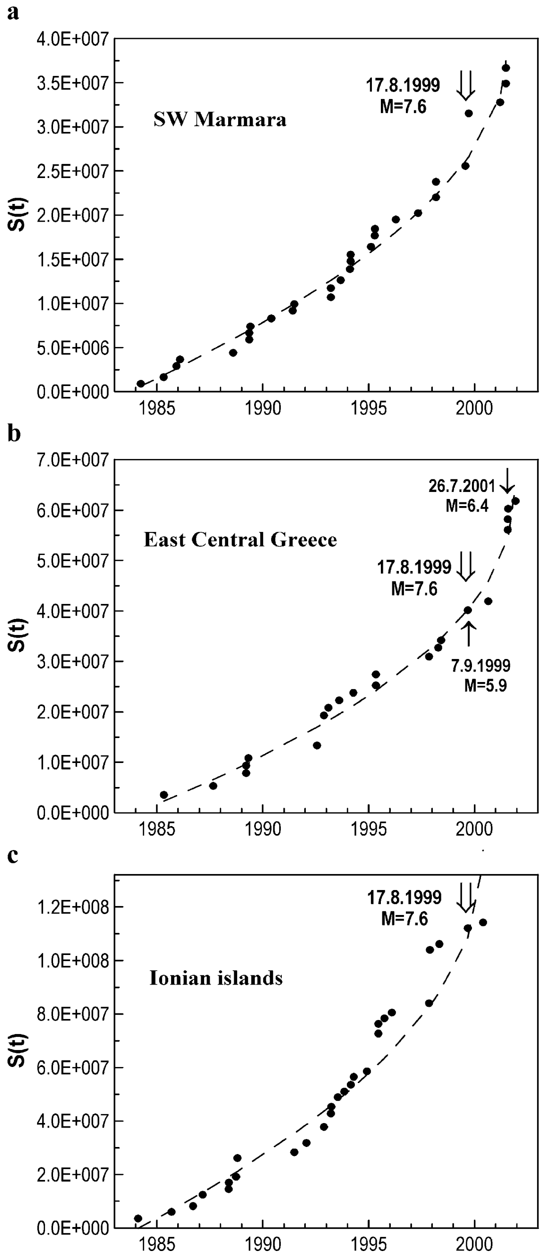

Fig. 2. Time variation of the cumulative Benioff strain, S, in the three critical regions for which information is given on Table 1. Large arrows indicate the Izmit 1999 mainshock origin time, whereas the small arrows in (b) correspond to the origin times if two strong shocks which occurred in the elliptical region of East-Central Greece (7.9.1999 M=5.9; 26.7.1002 M=6.4).

tion with time of the Benioff strain is obvious in all three cases. Moreover, it has to be noted that the acceleration of seismic deformation in the critical regions of SW Marmara (Fig. 2a) and East Central Greece (Fig. 2b) is more intense after the generation of the Izmit 1999 earthquake.

4. Discussion

Fig. 2a,b shows that the intense acceleration in the eastern and central part of the northern Aegean plate boundary started after the generation of the large Izmit 1999 earthquake. The largest up to now shocks in the eastern critical region (20.9.1999, $M=5.9$, 40.7°N, 27.6°E) and in the central critical region (26.7.2001, $M=6.4$, 39.1°N, 24.4°E) occurred also after the Izmit 1999 earthquake and their foci are in the northern Aegean plate boundary. This shows that the intense accelerating seismic excitation in the critical regions of the central and eastern part of the northern Aegean boundary has been triggered by the generation of the Izmit 1999 large earthquake. This is also supported by the fact that accelerating seismicity in these two critical regions could not be identified by the available data before the generation of the Izmit 1999 large earthquake.

Regarding the uncertainties in the identification of the basic focal parameters of the expected mainshocks (epicenter, magnitude, origin time), we adopted those mentioned previously (i.e. 120 km for the epicenter, ± 0.5 for the magnitude and ± 2 years for the origin time) and have been proposed by Papazachos et al. (2002b). They made such estimations on the basis of results of retrospectively predicted strong ($M \geq 6.4$) shallow mainshocks in the broader Aegean area. Moreover, Papazachos et al. (2002a), testing the efficiency of the proposed algorithm on synthetic but realistic random earthquake catalogs, found a possibility of 15% for misidendification of an area, which falsely exhibits an accelerated seismic deformation pattern compatible with the constraints imposed by relation (7).

Although the evidence for seismic excitation in the third region (C, Ionian islands) by the Izmit 1999 mainshock is not so strong, this critical region could not be identified before this mainshock. However, during the revision of the present paper, a strong mainshock ($M=6.4$, 14.8.2003, 38.76°N–20.60°E) occurred in the Ionian Sea, off the NW coast of Lefkada island, at a distance of about 70 km from the center of the respective critical region (entry 3 in Table 1). We believe that since the magnitude of this mainshock is smaller than the expected mainshock magnitude ($M=7.0 \pm 0.5$), further examination of this magnitude bias is necessary before this is declared as a successful prediction or not. Moreover, we note that the epicenters of almost all strong ($M \geq 5.8$) shallow earthquakes which occurred in the area of Greece after the 1999 Izmit mainshock were located along the North Aegean Boundary.

The triggering of accelerating seismic excitation along the northern boundary of the Aegean microplate by the generation of large mainshocks in the western part of the North Anatolian Fault supports the concept for a tectonic relation between the Anatolian microplate and the Aegean microplate (Papazachos, 2002). The fast westward coseismic motion of the Anatolia plate is expected to result in stress accumulation and trigger seismic slip along the North Aegean Boundary. This accumulation cannot be explained within the context of elastic stress increase due to the Marmara events, since static stress changes at more than a fault length from the source are negligible. On the other hand the increase in small-magnitude seismicity, which has been observed in the Aegean (west of 25°E) immediately after the passage of the Izmit mainshock surface waves in this area, was attributed to triggering resulted from the transient stresses caused by these waves (Brodsky et al., 2000). For this reason alternative models, possibly viscoelastic, might be used to explain such long-range interactions (Papazachos et al., 2000).

Information given in the present paper is not only of tectonic interest but of practical interest for intermediate term earthquake prediction and for time-dependent seismic hazard assessment. Independently of the generation or not of the large mainshocks for which information is given on Table 1, it is an observational fact (Fig. 2) that these three regions are in a state of accelerating seismicity due to the generation of intermediate magnitude earthquakes, some of which are strong enough ($M \sim 6$) and may cause considerable damage.

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