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Accelerating seismic crustal deformation in the North Aegean Trough, Greece

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SUMMARY

A recently developed algorithm has been applied to define regions of the northern Aegean in which accelerating seismic crustal deformation is currently occurring. An elliptical such region has been found in the western part of the North Aegean. Accelerating deformation, which started three decades ago and has been released by the generation of intermediate-magnitude earthquakes ($M \geq 4.5$), is still occurring. Based on these observations we can assume that this region is now in a state (pre-shock deformation) that will lead to a critical point (main shock).

The estimated basic parameters of this impending main shock are $\varphi = 39.7°$N, $\lambda = 23.7°$E for the epicentre, $M = 6.0$ for the moment magnitude, and $t_c = 2001.1$ for the origin time. The corresponding uncertainties are less than 100 km for the epicentre, ± 0.4 for the magnitude, and ± 1.5 yr for the origin time.

Key words: accelerating seismic deformation, Greece, North Aegean.

INTRODUCTION

Although the North Aegean area is far from the area in the Southern Aegean where the eastern Mediterranean lithosphere is being subducted under the Aegean (Papazachos & Comninakis 1969; McKenzie 1970, 1978; Le Pichon & Angelier 1979), it exhibits very high seismicity. The most prominent feature in this area is the North Aegean Trough (NAT), which represents the northwesternmost continuation of the North Anatolian Fault (NAF) in the North Aegean (Fig. 1), with the general earthquake pattern corresponding to dextral strike-slip faulting (Papazachos 1990). Indicative of the high seismicity of the area is the fact that, even over the last four decades, three very strong earthquakes have ruptured parts of the broader area of the NAT (Papazachos *et al.* 1999): the corresponding rupture zones are shown in Fig. 1 (1968 February 19, $M = 7.1$; 1982 January 18, $M = 7.0$; 1983 August 6, $M = 6.8$). There is also a long, well-documented, historical record of destructive earthquakes originating in the broader North Aegean area that have caused extensive damage in nearby areas (Papazachos & Papazachou 1997).

The main reason that motivated the present study is that, in a systematic search for accelerating seismic deformation caused by intermediate-sized events in the Aegean area, it was observed that the area along the North Aegean Trough is one of the areas where this accelerating deformation phenomenon currently occurs. An additional reason that justifies why this area merits particular attention is the evidence that strong earthquake seismic activity may be expected along the broader NAT region during the next few years due to triggering by the westward motion of the Anatolian plate following the recent large Izmit earthquake of 1999 August 17 ($M = 7.4$) (Papazachos *et al.* 2000a).

In the course of analysing seismicity as a random or critical phenomenon there has been an accumulation of evidence to suggest that some proportion of large and great earthquakes are preceded by a period of accelerating seismic activity of moderate-magnitude earthquakes that occur in the several years to decades prior to the occurrence of the large or great event, and over a region much larger than its rupture zones (Jaume & Sykes 1999). Early seismologists noted an increase in the occurrence rate of moderate-magnitude earthquakes before great earthquakes (Imamura 1937; Gutenberg & Richter 1954; Tocher 1959). Fedotov (1968) formulated the seismic cycle concept, and suggested an increasing seismicity rate before the second main shock of the cycle. Additional observations seem to favour this concept (Mogi 1977, 1981; Papadopoulos 1986; Scholz 1988, 1990; Karakaisis *et al.* 1991).

One of the main issues on which research on seismicity patterns has been focused during the last decade is whether the crust is in a continuous state of self-organized criticality or whether it repeatedly approaches and retreats from a critical state (Sammis & Smith 1999). Within this context, studies on the behaviour of intermediate-magnitude seismic activity prior to large earthquakes, mainly in California, resulted in the identification of accelerating seismicity, expressed in terms of seismic moment, energy or Benioff strain release before these earthquakes, which follows a power law (Varnes 1989; Sykes &

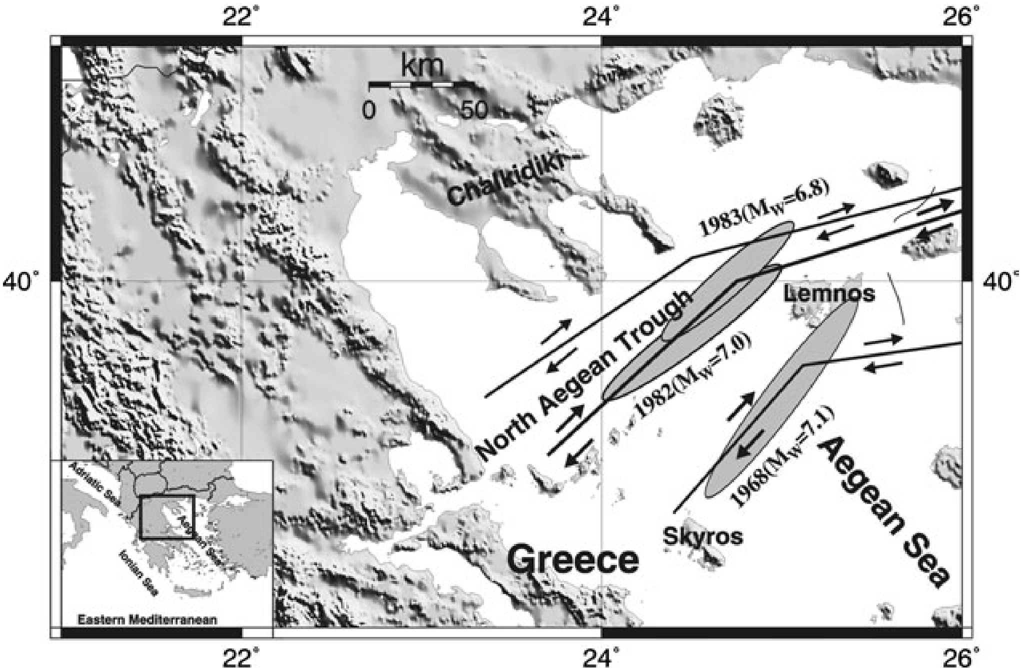

Figure 1. Map of the broader North Aegean area studied in this paper. The investigated area is delineated by a rectangle in the inset larger-scale map. The rupture zones of the 1968 ($M=7.1$), 1982 ($M=7.0$) and 1983 ($M=6.8$) main shocks are shown.

Jaume 1990; Bufe & Varnes 1993; Jaume & Sykes 1999). In these studies, the generation of intermediate-magnitude earthquakes (pre-shocks) was considered as a phenomenon culminating in a critical point, that is, the main shock, according to previous formulations (Sornette & Sornette 1990; Sornette & Sammis 1995; Huang *et al.* 1998).

During the last two years, a systematic study has been carried out to identify such phenomena in the broader Aegean area (Papazachos & Papazachos 2000a,b). The purpose of the present study is to identify the existence of accelerating seismic crustal deformation in the North Aegean area and to estimate the epicentre, magnitude and origin time of the impending event, assuming that this area is approaching a critical (pre-shock) state and its culmination.

METHOD AND DATA

Bufe & Varnes (1993), after studying the accelerating seismicity sequences in the San Francisco Bay area, proposed a power-law time-to-failure relationship for the changes in the rate of seismicity, expressed in terms of the cumulative Benioff strain, $S(t)$, that preceded the 1868, 1906 and 1989 large earthquakes. They defined $S(t)$ as

$$S(t)=\sum_{i=1}^{n(t)} E_i^{1/2}\,, \tag{1}$$

where E_i is the seismic energy of the ith pre-shock and $n(t)$ is the number of events at time t, and used a time-to-failure function of the following form to account for the time variation of $S(t)$:

$$S(t)=A+B(t_c-t)^m\,, \tag{2}$$

where t_c is the origin time of the main shock, and A, B, m are parameters that can be calculated with the available observations.

Bowman *et al.* (1998), aiming to test for the occurrence of accelerating Benioff strain release before large earthquakes in California, defined the critical region; that is, the circular region surrounding the main-shock epicentre in which the cumulative Benioff strain of the pre-main-shock seismicity best fits eq. (2). They proposed an optimization algorithm to quantify the accelerating crustal deformation (Benioff strain) by minimizing a curvature parameter C, defined as the ratio of the root mean square error of the power-law fit expressed by eq. (2) to the corresponding linear fit error. Hence, C is much less than 1 for accelerating or decelerating Benioff strain release and almost equal to 1 for a steady (linear) time variation.

Papazachos & Papazachos (2000a,b, 2001) studied the behaviour of the pre-shocks of strong earthquakes that occurred in the Aegean area and developed this methodology further by considering optimal elliptical regions where accelerating deformation can be identified. Moreover, they defined several relations that can be used as additional constraints for the model expressed by eq. (2), thus enhancing its robustness.

The following three relations hold between the magnitude, M, of the main shock, the radius, R (in km), of the circle with area equal to the area of the elliptical critical region, the parameter B of relation (2), and the average magnitude M_{13} of the three largest pre-shocks (σ is the standard deviation):

$$\log R=0.41M-0.64,\ \sigma=0.05\,, \tag{3}$$

$$\log B=0.64M+3.27,\ \sigma=0.16\,, \tag{4}$$

$$M=0.85M_{13}+1.52,\ \sigma=0.21\,. \tag{5}$$

For the duration, t_p (in years), of the pre-shock sequence, the following two constraints hold:

$$\log t_p=5.81-0.75\log S_r,\ \sigma=0.17\,, \tag{6}$$

$$A=S_r t_p\,, \tag{7}$$

where S_r (in $J^{1/2}yr^{-1}$) is the long-term Benioff strain rate within the elliptical critical region, and s_r is the same quantity reduced to 10 000 km^2.

The values of the parameter m of the relation (2) and of the curvature parameter C must be smaller than 0.7. That is,

$$m < 0.7,\ C < 0.7\,, \tag{8}$$

and their average values calculated for earthquakes in the Aegean area are $\bar{C}=0.46\pm0.13$, $\bar{m}=0.45\pm0.12$. Small C-values imply that the power law (eq. 2) fits the data better than the linear fit, while m-values larger than 0.7 lead to a time variation practically indistinguishable from the linear variation.

A parameter P has been defined as a measure of compatibility for the parameters R, B, M, t_p, and A, which are calculated for a given pre-shock sequence and the corresponding values determined for past earthquakes by eqs (3) to (7). P is the average of probabilities calculated for each of the left-hand-side parameters of these equations, assuming that the observed deviations for each parameter follow a Gaussian distribution. An additional quantitative measure of compatibility of the behaviour of a pre-shock sequence with that of the model expressed by the same relations is the ratio P/C. A study aiming at the *a posteriori* 'predicting' of past main shocks in the broader Aegean area showed that acceptable values of P are larger than 0.25 and those of P/C are larger than 0.36 (Papazachos & Papazachos 2001a); that is,

$$P > 0.25,\ P/C > 0.36\,. \tag{9}$$

The mean value of P is 0.40 ± 0.11 and that of P/C is 1.00 ± 0.46.

The accelerating deformation that satisfies all the above equations cannot be identified until a time t_i before the main-shock occurrence, which is the identification time for this critical phenomenon. Papazachos *et al.* (2001a) found that the difference $\Delta t_{ci}=t_c-t_i$ between the identification time and the origin time of the main shock is of the order of several years, and about 16 per cent of the duration of the pre-shock sequence:

$$\log(t_c - t_i) = 5.04 - 0.75\log s_r,\ \sigma = 0.18\,, \tag{10}$$

$$t_c - t_i = (0.16\pm0.05)t_p\,. \tag{11}$$

Identical results are reported by Yang *et al.* (2001), who independently found that only in the final 17 per cent of the pre-shock time period can the pre-shock region be identified. Hence, it is evident that the constraints expressed by eqs (3) to (8) do not allow the identification of accelerating Benioff strain release until a time close to the main shock, when this phenomenon is more pronounced. It is worth noting that eqs (10) and (11) can be used to estimate the main-shock origin time, since s_r and t_p can be calculated from the available data on the identification time.

According to the methodology applied (Papazachos & Papazachos 2000a,b), we first consider the earthquakes (pre-shocks) with epicentres in an elliptical region centred at a given point (observed or assumed main-shock epicentre), and then calculate the parameters of eq. (2) and the curvature parameter C. Calculations are repeated for a large set of geometrical characteristics of the elliptical area (varying length, a, azimuth, z, of the large axis of the ellipse, and ellipticity, e), various durations, t_p, of the pre-shock sequence, various main-shock magnitudes (ranging between $M=5.8$ and M_{max}, the maximum magnitude anticipated in each area), and several origin times of the impending main shock. All these calculations are repeated over a grid of points that covers the investigated area.

From all solutions obtained for all grid points, the one that fulfils relationships (3) to (9) and exhibits the smallest curvature parameter with the m parameter being low is chosen as the best solution, with the corresponding grid point, magnitude and origin time to be considered as the first approximations of the basic focal parameters of the impending main shock. The parameters that are finally adopted come from a second step of calculations that gives more satisfactory results, especially for the origin time. With the above-described procedure the origin time is estimated only very roughly, because it is very sensitive to errors introduced in the calculations. For this reason, another technique that gives a better estimate of the origin time has been applied. This technique is based on a change in the relation $T_i=f(T_c)$ between the assumed origin time, T_c, and the corresponding calculated identification time, T_i, when T_c becomes equal to the origin time t_c (Papazachos *et al.* 2001a). This change is usually an abrupt increase of T_i at the time $T_c=t_c$. This precursory phenomenon makes it possible to estimate the origin time with an error of less than one and a half years (±1.5 yr). Having a more accurate origin time we then repeat the calculation (c, etc.) for a more accurate epicentre and magnitude. The finally adopted epicentre coordinates are the average of the coordinates for the subgroup of solutions that have curvature parameters between C_{min} and $1.15C_{min}$ (assuming a 15 per cent error in C_{min}), and the finally adopted best solution is the one of this group of solutions for which the parameter m has the smallest value. The finally adopted magnitude for the ensuing main shock is the average of the three values given by relations (3), (4) and (5) for the best solution.

The data used in the present study have been taken from the catalogue of Papazachos *et al.* (2000b) and belong to one of the following three complete sets of shocks: 1911–49, $M\geq5.0$; 1950–64, $M\geq4.5$; 1965–2000, $M\geq4.3$. The errors in the epicentres are of the order of 15 km for earthquakes occurring after 1965 (when the first network of seismic stations was established in Greece) and up to 25 km for older earthquakes. All magnitudes are equivalent moment magnitudes and their errors are up to 0.3. The formula

$$\log E = 1.5M + 4.7 \tag{12}$$

derived by Papazachos & Papazachos (2000a) is used to calculate the seismic energy, E (in joules), from the moment magnitude, and then the cumulative Benioff strain is calculated by relation (1).

RESULTS

The broader North Aegean Trough area was searched using a grid of points 0.25° apart. For each point, C-values were calculated, according to the procedure described above, and the best solution was selected. Fig. 2 shows the spatial distribution of these C-values for m-values smaller than 0.6. An area of very low C-values can be observed in the western part of the NAT, while the grid point for which the C parameter exhibits its smallest value is located south of Chalkidiki. The above-described procedure was followed in order to determine

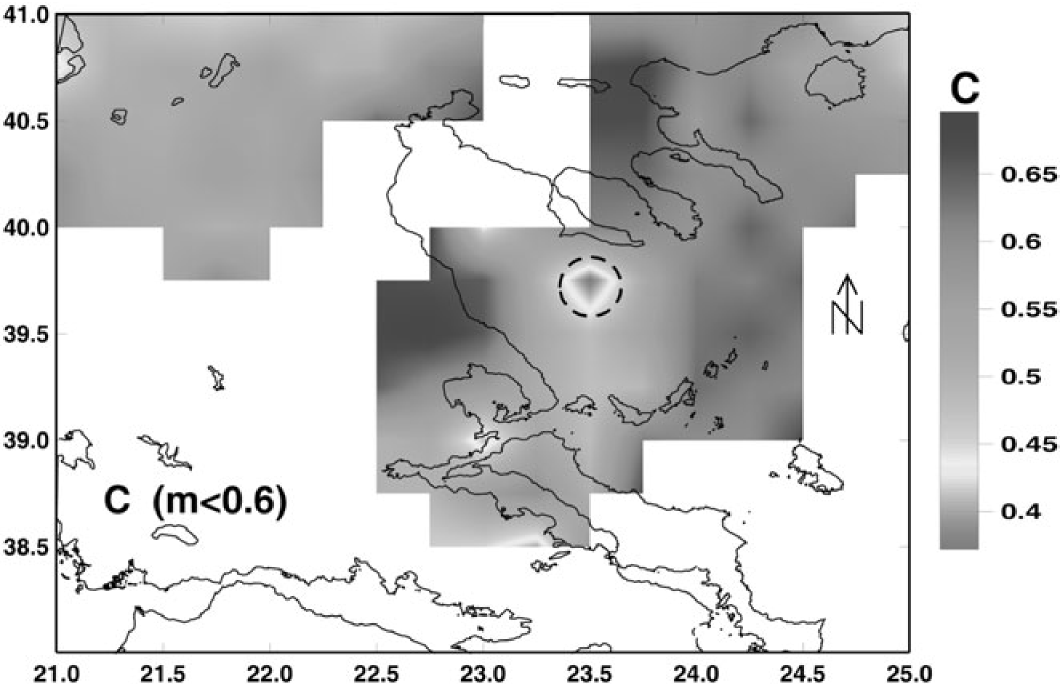

Figure 2. Curves of equal C-values in the broader North Aegean Trough (NAT) area for m smaller than 0.6.

the best solution and the parameters of the ensuing main shock. Table 1 summarizes these parameters; that is, the estimated epicentre coordinates of the expected main shock, its magnitude M and origin time t_c, and the model parameters (C, m, P/C, a, z, e). The minimum magnitude, M_{min}, of the pre-shocks considered, the number of observations, n (pre-shocks and main shock), and the year, t_s, when the accelerating pre-shock sequence started are also shown.

The critical region, where accelerating seismic activity is now observed, is shown in Fig. 3, along with the epicentres of the

Table 1. Epicentre coordinates (ϕ°_N, λ°_E), magnitude, M, and origin time, t_c, of the expected main shock, and parameters of the best solution.

ϕ°_N, λ°_E	M	t_c	C	m	P/C	a (km)	z	e	M_{min}	n	t_s
39.7°N, 23.7°E	6.0	2001.1	0.43	0.38	0.94	75	150°	0.90	4.5	23	1973

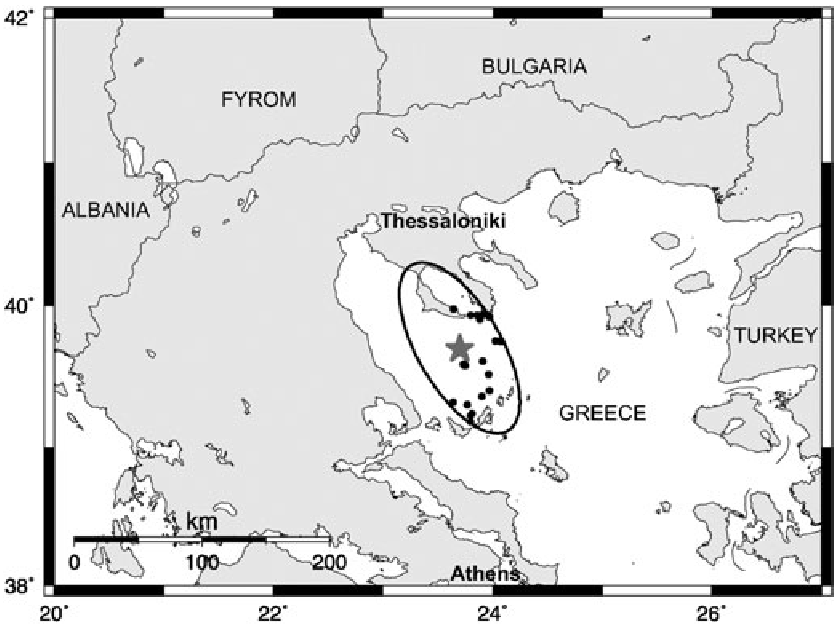

Figure 3. The elliptical critical region and the epicentres of the pre-shocks. The estimated epicentre of the expected earthquake is denoted by a star.

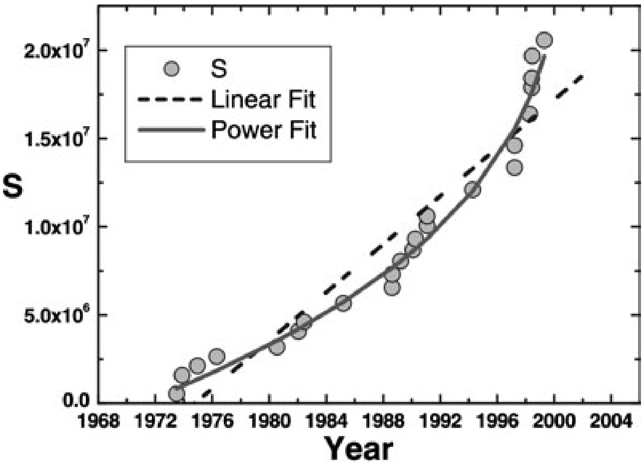

Figure 4. Time variation of the cumulative Benioff strain, $S(t)$, for the case investigated in this paper. The solid curve is the fit of the relation (2), and the dashed line is the linear fit.

pre-shocks from which data were used to calculate the Benioff strain S. The star denotes the proposed epicentre of the impending main shock. It is evident that the critical region, defined for the expected main shock, covers the low-C area shown in Fig. 2.

From Fig. 4, which is a plot of the cumulative Benioff strain, calculated using the available pre-shocks, against time, it is clear that the strain time variation is significantly accelerating. Fig. 5 shows the time variation of the C parameter, which was calculated according to the following procedure, without taking into account the impending main shock: the first six pre-shocks that occurred in the critical region were considered, and all parameters of eq. (2) as well as the curvature parameter C were determined using the value of t_c (main-shock origin time) already obtained. The number of pre-shocks was gradually increased until all pre-shocks were included. It is observed that the C-value continuously decreases as the pre-shock sequence approaches its end; that is, as the main shock is approached. A similar C-value behaviour has been found to precede the strong main shocks that have actually occurred in the Aegean area (Karakaisis *et al.* 2002).

A further parameter that has been found to qualitatively signify that the accelerating seismic activity in the critical region is related to an impending main shock is the b parameter of

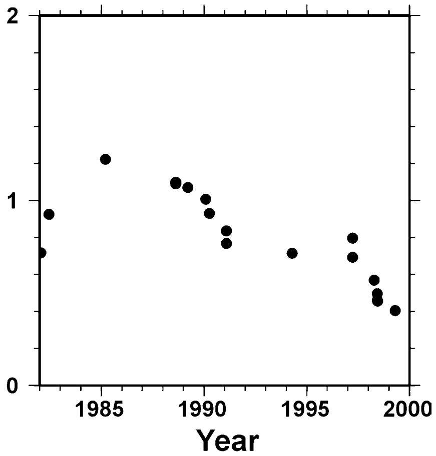

Figure 5. Time variation of the C-values (see text for explanation).

the Gutenberg–Richter recurrence law (Gutenberg & Richter 1944). Rundle *et al.* (2000) made the hypothesis that the characteristic earthquake is a first-order transition, and determined the cumulative magnitude distribution of earthquakes for three consecutive time periods during the earthquake cycle T, namely $0.8T$, $0.1T$ and $0.1T$. Small earthquakes occur uniformly at all times, but early on in the seismic cycle there is a systematic lack of intermediate-magnitude events, implying a high b-value. Then, during the following time period $0.1T$ there is an increase in the number of these events, which leads to a b-value lower than that in the previous period. Finally, during the last time period $0.1T$ there is a seismic activation and a further increase in the number of these intermediate-sized events, which results in an even lower b-value.

We tried to test the hypothesis of Rundle *et al.* (2000) for the earthquakes that occurred within the elliptical critical region shown in Fig. 3 during the pre-shock period. However, for statistical reasons and because we do not know what part of the seismic cycle corresponds to the pre-shock period, we did not fix the length of the time bins but instead used bins containing the same number of events. The need for a complete data set that is as large as possible led us to select all earthquakes that occurred within the critical region during the time period 1981–2000 with $M \geq 3.5$, because a complete data set is available for this period (Scordilis 1985). This data set was divided into three consecutive subsets, s_1, s_2, s_3, each containing the same number of earthquakes, i.e. 136 events.

The cumulative magnitude distribution of the earthquakes of these three subsets is shown in Fig. 6. It is evident that the b-value decreases from 1.79 in subset s_1 to 1.07 in subset s_3. Moreover, the shapes of the distributions are almost identical to those shown in Fig. 4 of Rundle *et al.* (2000, p. 2172).

In a different approach, and using all the earthquakes mentioned above (i.e. those with $M \geq 3.5$ occurring since 1981),

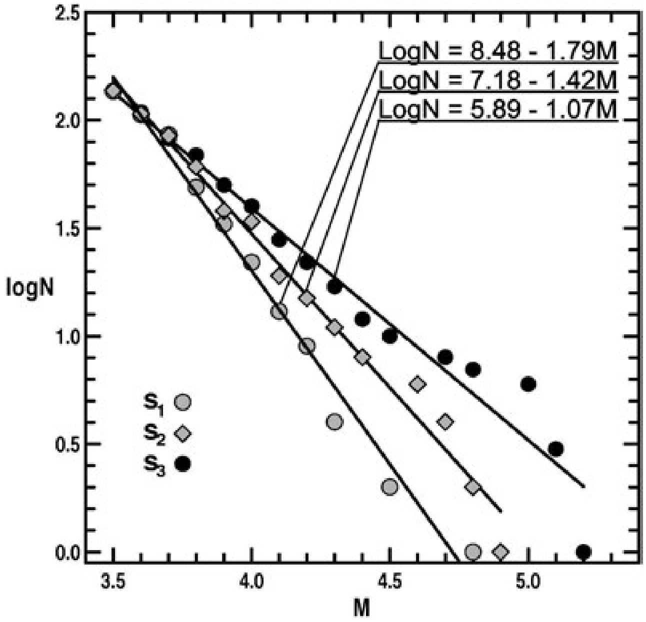

Figure 6. The cumulative magnitude distribution of the earthquakes with $M \geq 3.5$ that have occurred within the critical region of Fig. 3 since 1981. The whole data set was divided into three consecutive subsets, s_1, s_2, s_3, each containing the same number of events. The observed pattern is almost identical to the pattern predicted by Rundle *et al.* (2000).

consecutive b-values have been calculated for a moving window of at least 40 pre-shocks, with a step of one pre-shock, under the constraint that the magnitude range, ΔM, was at least 1.5 (Papazachos 1974). To normalize the variation of the b-values, we divided them by the b-value calculated using all the $M \geq 3.5$ events since 1981. These normalized b-values are plotted against the middle of the time interval they spanned and are shown in Fig. 7, along with their estimated errors. A decrease of the b-values can easily be seen as the end of the pre-shock sequence is approached, in agreement with recent relevant studies for main shocks that have already occurred (Karakaisis *et al.* 2002).

A crucial point of earthquake prediction is the uncertainties given for the basic focal parameters of the impending main shock. From a recent study concerning the retrospective prediction of several strong earthquakes that occurred in the broader Aegean area (Papazachos *et al.* 2001b), it has been deduced that the errors introduced by this method with a high probability (larger than 90 per cent) are 100 km for the epicentre, ± 0.4 for the magnitude, and ± 1.5 yr for the origin time. These uncertainties and probability level can also be adopted for the parameters listed in Table 1 for the impending main shock in the western part of the broader North Aegean Trough. It should be noted that the probability of occurrence of an event within the specified space–time–magnitude range by random chance is only 11 per cent. It is clear, however, that the retrospective analysis on its own is not enough to fully resolve the uncertainties involved in the prediction. Tests that are currently underway on synthetic random catalogues (Papazachos *et al.* 2001b), based on an approach similar to the one proposed by Zoller *et al.* (2001), suggest a 5 per cent probability for identifying an area that falsely exhibits accelerating seismic crustal deformation; that is, an area approaching its critical point although the seismicity within that area is assumed to be random in time and space.

DISCUSSION

There have been many approaches to the problem of earthquake prediction. In some of them it is supposed that the crust is always in a state of self-organized criticality (Bak & Tang 1989) and therefore earthquake prediction is inherently impossible (Geller *et al.* 1997). In the last decade, however, a growing number of observations, supported by results obtained through models of simple cellular automatons with loss and/or structural complexity, have suggested that a large earthquake perturbs the surrounding region away from the critical state and that methods of statistical physics can be used to monitor the return of the region toward criticality and the next large earthquake (Sykes & Jaume 1990; Triep & Sykes 1997; Sammis & Smith 1999).

A growing number of observations has shown that there are only a few cases for which strong earthquakes were not preceded by accelerating seismic deformation (Bowman 1992; Bufe *et al.* 1994), although it is not known if this activity always leads to a main shock. In most cases, however, an accelerating occurrence rate of moderate-magnitude seismicity has been observed before strong main shocks. This behaviour is more pronounced and fairly recognizable if expressed in terms of Benioff strain release (Bufe & Varnes 1993).

Detailed studies on seismicity patterns of the broader Aegean area show that all strong main shocks that have occurred during the last 50 years (1950–2000 with $M \geq 7.0$; 1980–2000 with $6.9 \geq M \geq 6.4$) were preceded by a detectable accelerating seismic activity (Papazachos & Papazachos 2001b). Considering these results and taking into account the fact that the parameters P and P/C, which express the compatibility of the behaviour of the currently evolving pre-shock sequence in the broader North Aegean Trough area studied in the present paper with the average behaviour of observed pre-shock sequences, have quite large values, it can be concluded that the generation of a strong main shock in the western part of the NAT seems quite probable.

It has to be borne in mind that the broader NAT area is a part of the North Aegean Boundary (which terminates at Cephalonia island, Ionian Sea), and that independently obtained results concerning this boundary suggest high probabilities for the generation of strong main shocks along this boundary during the three years following the Izmit, NW Turkey, earthquake of 1999 August 17 ($M = 7.4$) (Papazachos *et al.* 2000a).

To summarize, we suggest that the western part of the North Aegean Trough, south of Chalkidiki peninsula, has been exhibiting accelerating seismic activity since 1973, and that there is strong evidence that this area is approaching a critical

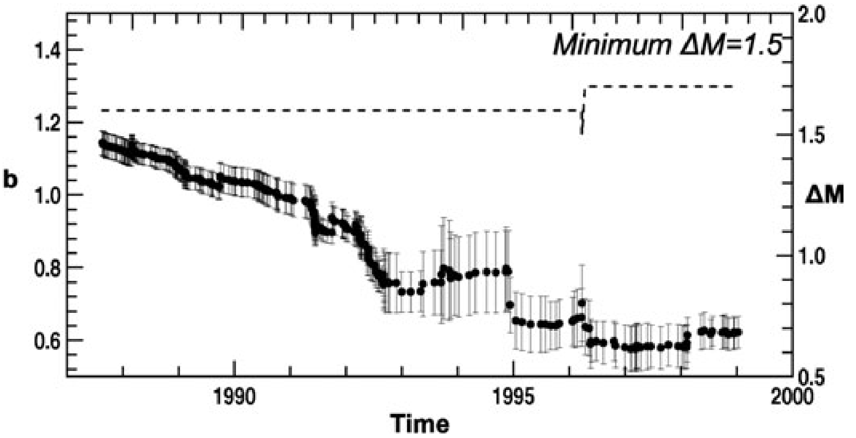

Figure 7. Time variation of the b-values. The dashed line shows the magnitude range, ΔM, that applies for the calculations (see text for explanation). Error bars are also shown.

state, expected to culminate in a strong main shock with a magnitude of about 6.0 before the end of 2002. This result, combined with independent evidence for strong earthquake seismic activity in this area during the next few years, suggest that measures aiming at mitigating the consequences of the impending main shock should be taken.

During the revision of the present work, a strong event ($M_w = 6.3$) occurred on 2001 July 26 with epicentre coordinates $\varphi = 39.05°$N, $\lambda = 24.35°$E, approximately 90 km from the proposed epicentre. Although the event occurred on the southern (Skyros) branch of the North Aegean Fault system, the epicentre, origin time and magnitude of this earthquake are within the error limits defined in this paper, and hence it could be considered as a successful forecasting based on the method applied here.

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