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Did the 7/9/1999 M5.9 Athens Earthquake Come with a Warning?

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Abstract. Prior to the 7/9/1999 $M_S = 5.9$ Athens earthquake, regional seismicity has exhibited a power-law increase, of the form $\Sigma\Omega = K + A(t_c - t)^n$, where Ω is estimated using an expression log $\Omega = cM + d$ and t_c is the time of the culminating event. Such changes appeared after the 17/8/1999 M7.4 Izmit event. We quantified the performance of the power law vs. the null hypothesis of constant seismic release rates, by defining the curvature C as the ratio of the power law fit RMS/linear fit RMS, so that the smaller C is, the better the power law behaviour. By mapping C, we have established a critical radius of 110 km and observed that the region of correlated accelerating seismic release extended from the N. Aegean, through Euboea and Attica to the SW Peloponnese. A few days prior to the Athens event, min(C) was centred at the epicentral area and numerical simulation yielded $t_c = 1999.676$ and predicted M_S = 5.77. Seismicity rates returned to normal (quasi-constant) after the Athens event. We interpret this effect as critical point behaviour, following remote excitation of a broad area by stress redistribution due to the Izmit event which, at Athens, has triggered 'premature' failure of a fault nearing its load bearing capacity. If this is correct, we have documented a case of remote earthquake triggering by another earthquake, as well as insight into the mechanisms producing it. As a corollary, we note that a large event may beget another large event in its broader region of interaction, which may be preceded by characteristic precursory seismicity changes.

1. Introduction

The damaging 7/9/1999 M = 5.9 Athens earthquake came rather unexpectedly in an area thought to have reduced earthquake hazard, although some early warning of its seismogenetic potential was given as early as 1956 (see Galanopoulos, 1956 and references therein). As if this wasn't enough, it arrived very shortly after the major 17/8/1999 M_W = 7.4 earthquake of Izmit, Turkey, which occurred approximately 650 km to the NE of Athens. This raises the natural question of whether the two events are causally related and if so, how. In addressing such questions, one has to deal with the complex and as yet unsolved problems of long range interactions in constrained, multiply correlated, many fault systems.

The problem of fault interactions and earthquakes triggering other earthquakes is addressed from many different viewpoints. Statistical mechanics are used ana-

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lytically (e.g. Czechowski, 1991; Rundle *et al.*, 1995), or by means of simulations such as slider-block models (e.g. Huang and Turcotte, 1990; Morein *et al.*, 1997) and cellular automata (e.g. Bak and Tang, 1989). Other authors prefer to simulate complex interacting systems (e.g. Senatorski, 1997) or treat physical models of faults in an elastic medium by investigating the possible effects of static and/or transient stress transfer due to slipping faults (e.g. see Harris, 1998 for a review and references therein for details). The emerging view of seismogenesis as a Self-Organised Critical (SOC) process in a complex system introduces new tools in the analysis of fault interactions, with methods drawn from statistical mechanics and many-body physics. During the recent several years, a mounting body of evidence indicates that the earthquake generation process can be viewed as a critical phenomenon culminating with a large event that corresponds to some critical point (Allègre and LeMouël, 1994; Keilis-Borok, 1990; Saleur *et al.*, 1996a, b; Sornette and Sammis, 1995; Bowman *et al.*, 1998). It has also been demonstrated that rupture in heterogeneous media is a critical phenomenon (Herrmann and Roux, 1990; Vanneste and Sornette, 1992; Sornette *et al.*, 1992). According to the Critical Point (CP) hypothesis, failure in the crust can be thought of as a scaling up process, in which failure at one scale is part of the damage accumulation over a larger scale, leading to long range stress-stress correlation and an increase (acceleration) of seismic activity prior to large earthquakes. This may be described by a power law time-to-failure relation of the form

$$\frac{\mathrm{d}\sum\Omega(t)}{\mathrm{d}t} = k(t_c - t)^{-a} \Rightarrow \sum\Omega(t) = K + A(t_c - t)^n \tag{1}$$

where Ω is any quantity estimated from the earthquake magnitude using an expression of the form $\log\Omega = cM + d$, (including earthquake count when $c = 0$), $\Sigma\Omega$ is the cumulative seismic release as a function of time, t_c is the time at which a critical state is attained, A is negative and $n < 1$, of the order of 0.3. K is the value of $\Sigma\Omega$ at $t = t_c$ and is finite, allowing the estimation of magnitude when $\Sigma\Omega$ represents some form of energy or moment release. This scaling law has been justified in terms of run-away crack propagation and empirical expressions for accelerating (tertiary) creep preceding failure in the laboratory (Voight, 1989; Varnes, 1989; Bufe and Varnes, 1993), but it can also result naturally from the many- body interactions between small cracks forming before the impending rupture (Sornette and Sammis, 1995; Saleur and Sornette, 1996; Bowman *et al.*, 1998). In this case, small and intermediate size events are associated with the increasing correlation length of the regional stress, while the culminating earthquake in the cycle represents the critical point occurring when the system is correlated over long ranges.

The critical state and the postulated self-organised criticality of the earth's crust (Bak and Tang, 1989; Scholtz, 1991; Sornette, 1991; Main, 1997) describe properties of the crust at different time scales. As Huang *et al.* (1998) indicate, the critical nature of large events results from the interplay between the long-range stress-stress correlations of the self-organised critical state and the hierarchical structure in such a way, that hierarchical rupture at a given level is like a critical point to the lower levels. Critical rupture occurs when the applied force reaches a value beyond which the fault system moves globally and abruptly, while self-organised criticality needs a slow driving force and describes the jerky steady state of the system at large time scales: the critical state of large earthquakes is only a fraction of the long-term condition described by self-organised criticality (Bowman *et al.*, 1998). However, and although there exist many examples of precursory seismicity changes fitting the CP earthquake model, it is not as yet clear whether all earthquakes behave as critical point processes. Moreover, as Bowman *et al.* (1998) point out, in a CP system without a single largest fault element to rupture, the long range correlation of the stress field may either allow the final event to connect neighbouring faults or, may simply remove the system from criticality with a number of small(er) earthquakes. Such topics are still under investigation and much effort is needed before definite answers exist.

Seismicity is the primary source of data for studying the phenomenology of fault interactions and constraining physical models. For instance, it has been attempted to demonstrate by means of statistics, that some areas may be correlated in such a way, that earthquakes occurring in one region may trigger earthquakes in another region, in a repeatable fashion (e.g. Papadopoulos, 1997; Papadopoulos and Drakatos, 1998). The consensus from hitherto large scale observational studies is that both near and far field earthquake triggering is identified by periods of increased seismicity rates, but there's no definitive information as to the rise and decay modes of such changes (for instance Reasenberg and Simpson, 1992; Dieterich, 1994; Stein *et al.*, 1997; Toda *et al.*, 1998; Nalbant *et al.*, 1998). The situation is aggravated in the case of far field triggering, where limited observational data exists, with contradictory (e.g. Hill *et al.*, 1995; Jones and Haukson, 1997) or negative (e.g. Gomberg and Davis, 1996) results. Moreover, practically nothing is known about possible seismicity changes prior to some large event hastened through remote interaction by a great earthquake, as may be our case.

Of all fault interaction models that can be tested with seismicity data, the time-to-failure/CP hypothesis is most interesting because of the simple predictive expression (1). In marked contrast, the precursory quiescence hypothesis is not endowed with such a convenience. Although the concept is independent of scale, experimental verification of precursory accelerating seismicity has focused on treating long-term processes leading to large scale rupture (e.g. Varnes, 1989; Bufe and Varnes, 1993; Bufe *et al.*, 1994; Bowman *et al.*, 1998; Brehm and Braile, 1998, 1999). Consequently, the concept has only been reckoned as a tool of intermediate term earthquake prediction. It has not been considered as a model of interactions following a massive regional stress perturbation (e.g. due to a great earthquake), neither has it been evaluated as a short term prediction tool. Having in mind the shortage of rigorous observational constraints on fault interaction models and the questions pertaining to the applicability of the CP model in particular, we set out to investigate the seismicity and seismic release rates prior to the Athens event and its possible relationship to the great Izmit earthquake.

2. The Data

The seismicity data used in this study are taken from the raw catalogue of the Geodynamics Institute of the National Observatory of Athens (NOA, http://www.gein.noa.gr/services/cat.html) and span the period 1/10/1998–16/10/1999 (Figure 1). The NOA catalogue is remarkably homogeneous. Figure 2a shows the cumulative number of events and Figure 2b compares the seismicity rates between the randomly chosen periods $[t_1, t_2]$ and $[t_3, t_4]$ shown in Figure 2a, in terms of their frequency-magnitude distribution. As can be seen, the two rates are almost identical, meaning that no inadvertent changes in the number and magnitude reporting rates have taken place during the period under consideration. The threshold of completeness is approximately $M_L > 2.8$, (Figure 2a). However, noting that small distributed variations in the frequency-magnitude statistics do not warrant the use of earthquake count to evaluate regional rate changes, we shall only rely on measures of energy (small earthquakes exert little influence on the cumulative seismic energy release). Herein we use cumulative Benioff strain as a function of time, which is defined as

$$\epsilon(t) = \sum_{i=1}^{N(t)} \sqrt{E_i(t)}, \qquad (2)$$

where $E_i(t)$ the energy of the ith event and $N(t)$ is the total number of events at time t. Kanamori and Anderson (1975), provide the relationship

$$\log_{10} E_i(t) = 4.8 + 1.5M_S \qquad (3)$$

and M_S can be estimated from the M_L reported by NOA as

$$M_{S(Athens)} = M_L + 0.5. \qquad (4)$$

3. Data Analysis I: Identification of Accelerated Seismicity

Figure 3 shows the cumulative Benioff strain from 1 June–6 September 1999, computed over expanding circular areas concentric with respect to the Athens epicentre, using the data listed in Table I. The left vertical dotted line indicates the origin time of the 17/08/99 Izmit event and the right dotted line the same for the 7/9/99 Athens event. Power law acceleration of seismic release rates is apparent at all radii but the longest, and while it is rather subtle prior to the Izmit event, (slightly deviating from constant rate), it becomes prominent right afterward.

Table I. All earthquakes with hypocentres shallower that 60 km, located within a radius of 140 km from the NOA epicentre of the 7/9/1999 Athens earthquake (38.15°N, 23.62°E), during the period 10/5/1999–7/9/1999 (source, http://www.gein.noa.gr/services/cat.html)

Date	Time	Latitude (°)	Longitude (°)	Depth (km)	M_L
2/6/1999	19:33:23	38.16	22.16	35	3.0
5/6/1999	06:19:18	38.29	22.35	5	4.1
6/6/1999	05:14:37	39.14	23.71	28	3.1
7/6/1999	21:57:09	38.37	22.27	5	3.0
11/6/1999	01:12:00	37.69	23.85	5	2.3
11/6/1999	16:00:18	38.85	23.48	17	3.0
11/6/1999	23:02:36	38.68	24.84	46	3.2
20/6/1999	06:37:40	37.99	22.43	39	3.0
21/6/1999	05:37:43	39.34	23.14	5	3.4
25/6/1999	07:42:15	38.30	22.76	11	4.3
25/6/1999	19:39:46	37.90	23.28	29	2.7
25/6/1999	19:46:29	38.00	23.17	11	2.3
26/6/1999	00:42:35	38.17	23.36	58	2.4
27/6/1999	23:07:27	38.80	23.22	5	3.1
29/6/1999	15:10:00	38.40	22.05	5	4.4
3/7/1999	21:35:50	39.01	24.48	5	2.9
5/7/1999	07:46:29	38.53	22.19	9	3.1
7/7/1999	20:25:23	38.75	22.25	5	3.3
8/7/1999	06:16:35	38.39	22.05	5	3.0
9/7/1999	13:37:50	38.00	23.70	10	2.9
11/7/1999	11:01:38	37.29	24.72	9	3.2
20/7/1999	00:05:04	37.44	24.05	33	2.6
28/7/1999	07:14:05	38.14	22.05	36	3.2
29/7/1999	14:53:18	38.36	22.07	15	3.0
31/7/1999	07:19:11	38.38	22.05	5	3.6
31/7/1999	10:33:45	38.59	24.41	36	3.2
2/8/1999	08:46:46	38.34	22.35	10	2.7
3/8/1999	09:43:24	37.70	23.84	37	3.0
4/8/1999	01:36:31	38.93	22.57	35	3.1
4/8/1999	02:17:23	38.30	22.24	5	3.1
4/8/1999	05:00:14	37.06	23.63	5	2.9
4/8/1999	06:30:56	37.89	22.10	28	2.8
5/8/1999	17:59:29	38.35	22.27	5	3.0
6/8/1999	03:10:11	38.48	23.88	5	2.5
8/8/1999	12:13:38	38.12	23.35	5	2.9
10/8/1999	20:52:27	38.87	24.04	29	2.8
11/8/1999	06:02:55	38.04	22.14	12	3.0

Table 1. Continued

Date	Time	Latitude (°)	Longitude (°)	Depth (km)	M_L
18/8/1999	02:17:54	38.27	23.28	5	2.5
18/8/1999	19:14:31	38.36	25.20	5	3.7
20/8/1999	13:44:29	38.95	23.28	10	3.5
22/8/1999	19:36:56	38.15	22.65	10	2.6
23/8/1999	04:06:01	38.08	23.88	18	2.4
23/8/1999	06:38:12	38.03	23.93	17	2.5
27/8/1999	01:06:50	39.30	23.40	40	3.0
27/8/1999	01:14:20	39.24	23.48	3	3.0
27/8/1999	04:06:42	37.86	23.78	10	1.8
27/8/1999	15:39:06	38.05	22.58	5	3.6
27/8/1999	15:44:21	38.38	23.90	6	3.3
27/8/1999	20:08:42	38.12	23.88	18	2.4
28/8/1999	02:18:44	37.94	23.05	13	2.8
30/8/1999	22:14:26	38.92	23.16	26	3.0
3/9/1999	05:29:34	38.49	23.35	33	4.1
4/9/1999	00:07:20	38.04	22.40	5	3.0
4/9/1999	05:45:02	39.03	23.39	11	3.2

All curves of Figure 3 can be modelled with a power law relation of the form (1). We have implemented a Hedgehog non-linear optimisation procedure, minimising the L_2 norm. Moreover, following Bowman *et al.* (1998), we quantify the performance of the power law fit against the null hypothesis of constant seismic release rate by defining a curvature parameter

$$C = \text{(Power law fit RMS)/(Linear fit RMS)},$$

such, that when the data are best described by a power law curve, the RMS error will be small compared to the RMS error of the linear fit and C will also be small. An example of this concept is shown in Figure 4c.

We adopt a two stage modelling procedure. Since we make posterior analysis and some cardinal parameters as the epicentre and t_c are known, we can begin with constrained modelling that facilitates determination of other important factors, such as are the correlation length or critical radius R_c and the predictable magnitude. Apparently, the optimal R_c will be found at min(C), while by virtue of (2), Equation (1) reduces to $\epsilon(t \equiv t_c) = K$ and can be inverted to a predicted magnitude $\hat{M}_S$. Using the data of Table I for events above the magnitude of completeness, ($M_L \geq 2.8$ or $M_S \geq 3.3$), and by fixing t_c at 00:00 of 7/9/1999, we obtain the results of Figure 4a,b. The optimal radii R_c are found to be 100–120 km around the epicentre, (Figure 4a), and the corresponding $\hat{M}_S$ are found to vary

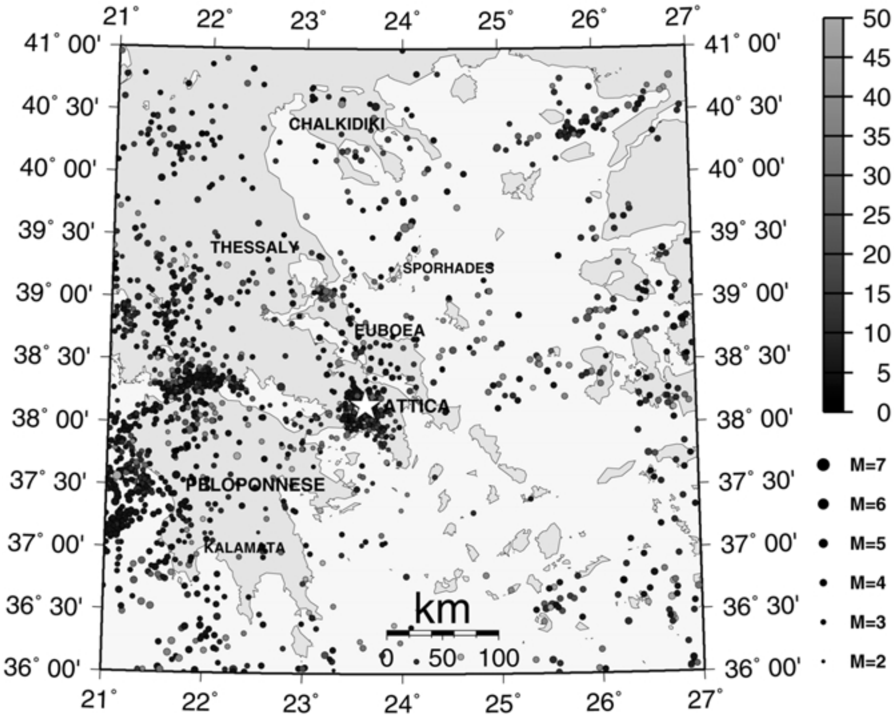

Figure 1. The catalogue of the Geodynamics Institute, National Observatory of Athens spanning the period 1/10/1998–16/10/1999. Magnitude scale is the M_L scale reported by NOA.

in the range 5.7–5.85, (Figure 4b), consistent with the observed M_S of 5.9. Having determined the optimal size of the critical radii, we let t_c vary in order to investigate its predictability. The results are shown in Figure 4c for R_c = 110 km. The dashed line indicates the best fitting straight line, which represents the null hypothesis of constant seismic release rates. The continuous line indicates the best fitting power law model. The predicted critical time $\hat{t}_c$ is determined at 2.76 days (66 hours) prior to the origin time of the main shock (indicated with a vertical bar). Finally, letting R_c = 110 km, we simulate a conditional running forecast of the Athens main shock, calculating C, $\hat{t}_c$, and $\hat{M}_S$ as a function of the time-to-failure for the period 16/8/99–7/9/99, in steps of one day. Figure 4d is illustrates the change in curvature C and demonstrates the evolution of seismic release, from weakly to fully developed power-law behaviour prior to the Athens event. Figures 4e and 4f, respectively illustrate the corresponding improvement in the determination of predicted critical time and magnitude, to their final values of $\hat{t}_c = -2.76$ days and $\hat{M}_S = 5.77$. In fact, $\hat{t}_c$ is almost the same as the time of the last event prior to the Athens main shock. No seismic activity is observed thereafter at the level of $M_S > 3.3$, and the entire region within 100–120 km from the epicentre went quiescent almost 66 hours prior to failure. This may imply that all stress-stress

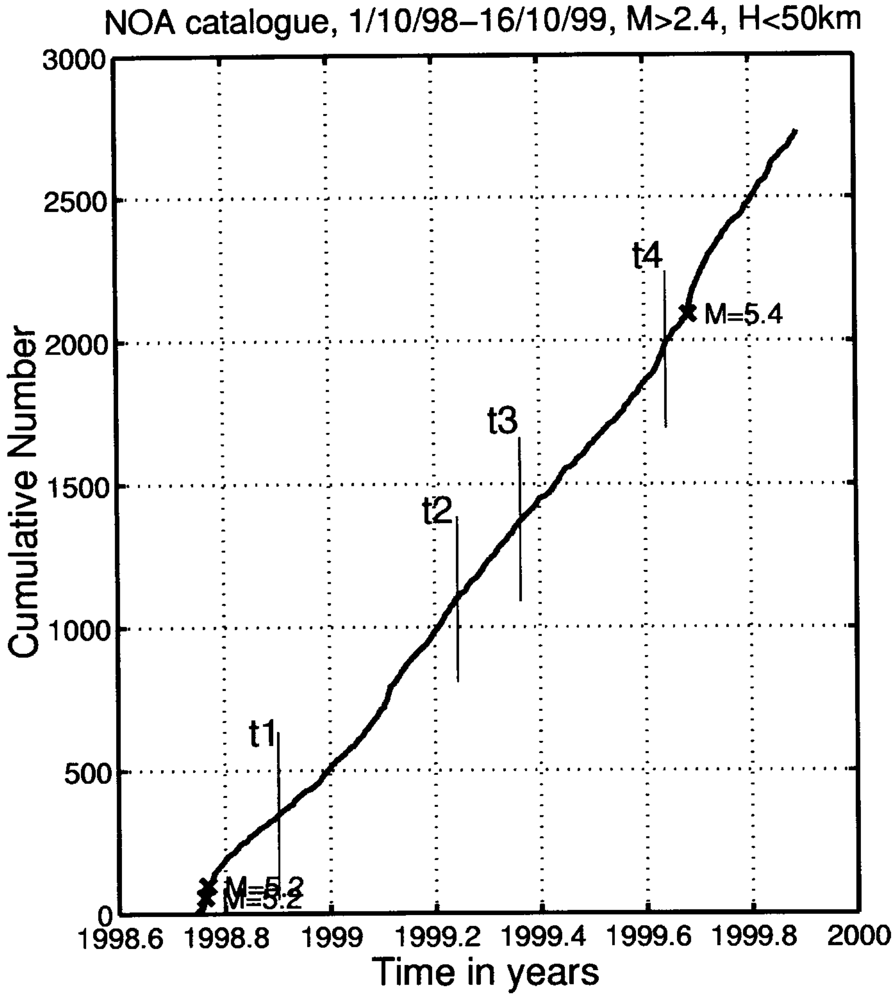

Figure 2a. Cumulative number of events for the NOA catalogue (1/10/1998–16/10/1999). t_1, t_2, t_3 and t_4 define the time intervals used for comparison of seismicity rates in Figure 2b.

correlations that could be made had been completed by that time and the regional system had entered the critical phase, only waiting for some additional factor to instigate irreversible instability (rupture).

4. Data Analysis II: Space-Time Dependence of the Acceleration Process

The significant acceleration of seismic release rates right after the 17/8/1999 Izmit earthquake hints at a causal connection between the two events and raises the issue of the spatial distribution and temporal behaviour of their correlation process. We have found that mapping of the spatial and temporal evolution of C (cf. Figure 4d) on a geographical grid can be a very useful tool in searching for the answer

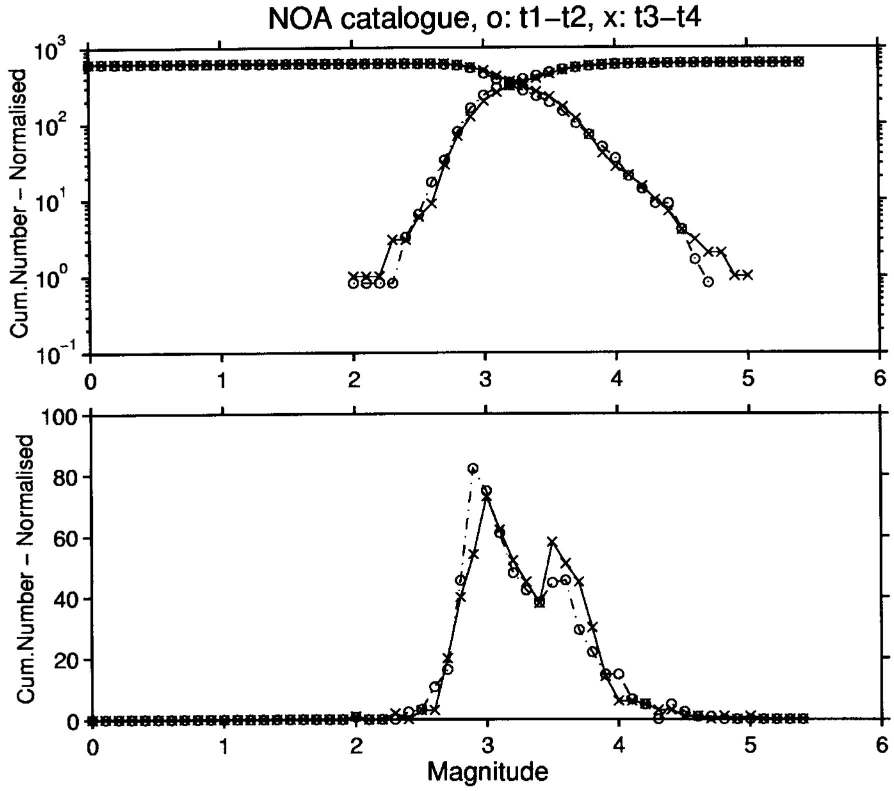

Figure 2b. Comparison of seismicity rates between the periods $[t_1, t_2]$ and $[t_3, t_4]$ shown in Figure 2a, in terms of their frequency-magnitude distribution (top) and number per magnitude (bottom).

to such questions. Accordingly, we present four curvature maps at different cut-off times from 16 August to 6 September 1999. The analysis included all events above the magnitude of completeness of the NOA catalogue (cf. Figure 2b, $M_L \geq 2.8$, $M_S \geq 3.3$). A constant correlation radius of 110 km is used throughout. The results can be summarised as follows.

Time cut at 16/8/1999 (1999.6246). Figure 5 illustrates the distribution of curvature computed over the period 15/6–16/8/1999, ending just prior to the Izmit event. Note that areas containing fewer than 5 events within the 110 km radius are left blank, because a reliable power law model is not guaranteed. Evidently, C is of the order of 1 or higher in the greatest part of the mapped area, meaning quasi-constant seismic release rates. There's one exception in the area of Kalamata, SW Peloponnese, where $C \approx 0.3$. The power law model of this data indicates an event of magnitude 5.8 to occur on, or right after 17/8/1999. This is a false alarm for that area, but an earthquake did occur on 17/8 albeit at Izmit, more than 800 km away. It is a puzzling question whether such power law behaviour at Kalamata is coincidental, or comprises the effect of long range correlation with the process of

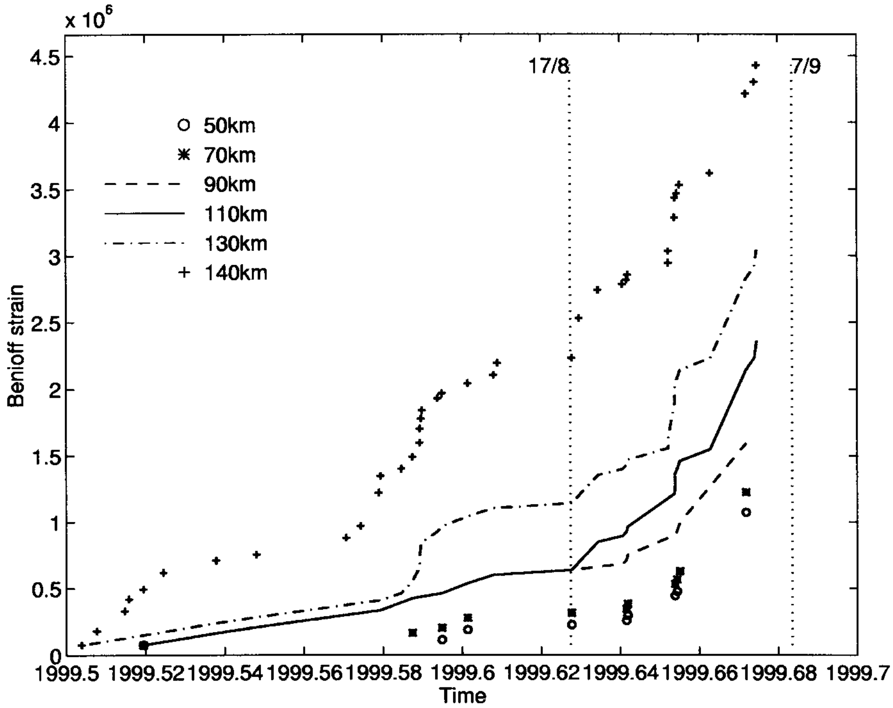

Figure 3. Cumulative Benioff strain from 1 June–6 September 1999, computed over concentric circles, centred on the epicentre of the Athens earthquake.

the Izmit event. An answer cannot be given at present but the data presented in the ensuing analysis is consistent with the latter hypothesis.

Time cut at 31/8/1999 (1999.6657). Figure 6 illustrates the distribution of curvature during 25/6–31/8/1999. The situation has changed dramatically during the time elapsed since the Izmit event, with clear areas of $C < 0.4$, (correlated seismic release), appearing at offshore Thessaly and the Sporhades islands, (at the western termination of the North Aegean Trough), Attica and central-east Peloponnese. These are surrounded by a halo of relatively low C values (<0.6).

Time cut at 6/9/1999 (1999.6822). Figure 7a corresponds to the period 25/6–6/9/1999, i.e. just prior to the Athens event. It is readily seen that the area of correlated seismic release includes all the regions of low C values depicted in Figure 7, and has also expanded to a more or less continuous belt from the Chalkidiki peninsula through offshore Thessaly and the Sporhades islands, Euboea and Attica, to the central-west Peloponnese. The lowest C is observed in Sporhades, Euboea and Attica, but the absolute minimum (0.18) is located precisely at the epicentral area of the Athens event. To ascertain that the results of Figure 7a are statistically significant, the *resolution*, i.e. the number of earthquakes used to calculate C at each grid point is shown in Figure 7b. It is evident that upwards of 20 events above

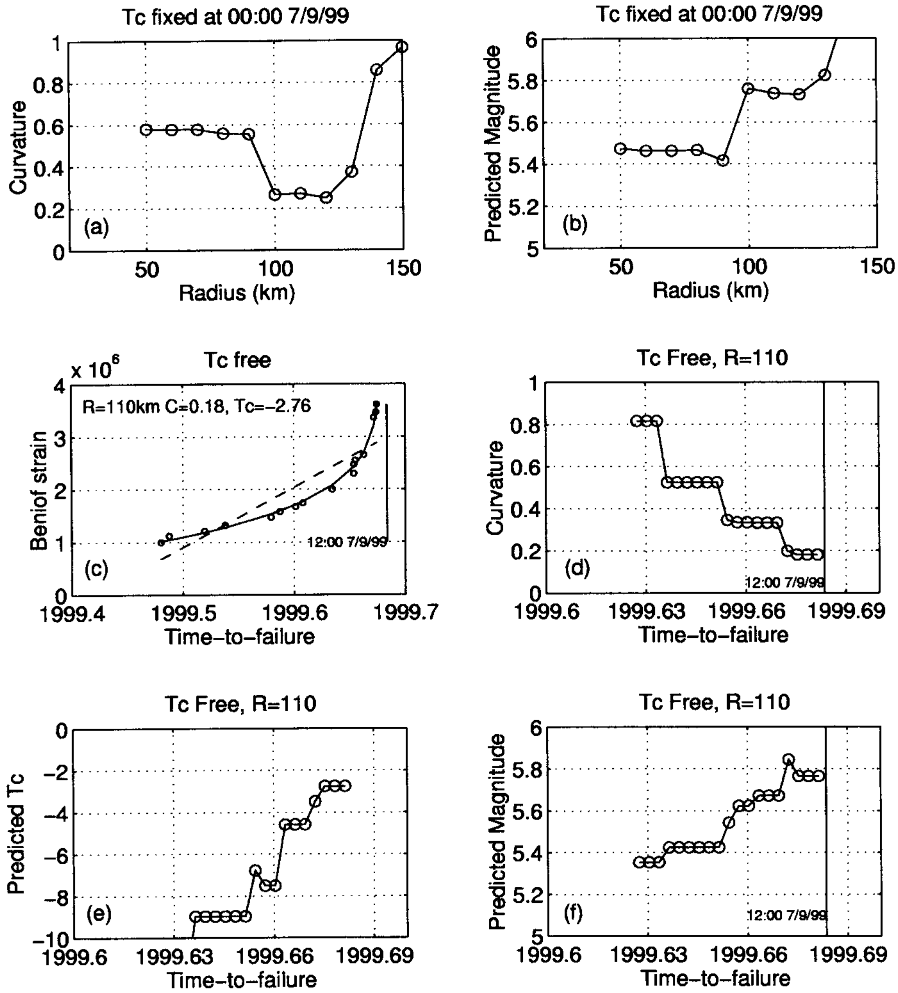

Figure 4. (a) The variation of curvature vs. distance from the epicentre for t_c fixed at 00:00 of 7/9/99. (b) The variation of predicted magnitude $\hat{M}_S$ vs distance from the epicentre for t_c fixed at 00:00 of 7/9/99. (c) Modelling of the cumulative Benioff strain within the optimal critical radius, leaving t_c unconstrained. In all drawings a–d, the data span the period 1 June–6 September 1999. (d) The variation of curvature vs. time-to-failure for t_c unconstrained. (e) The variation of predicted critical time vs. time-to-failure, for $\hat{t}_c$ unconstrained. (f) The variation of predicted magnitude $\hat{M}_S$ vs. time-to-failure for t_c unconstrained.

the threshold of completeness were involved in the calculations, in almost all areas where power-law acceleration was observed.

Time cut at 16/10/1999 (1999.7917). Figure 8a, illustrates the distribution of curvature for the period 1/8–16/10/1999, i.e. including the inter-event acceleration period and the post-Athens earthquake days. The analysis of this data set is complicated by the Athens aftershock sequence, which influences the computation of short-term regional seismicity rates. Aftershocks are thought to result

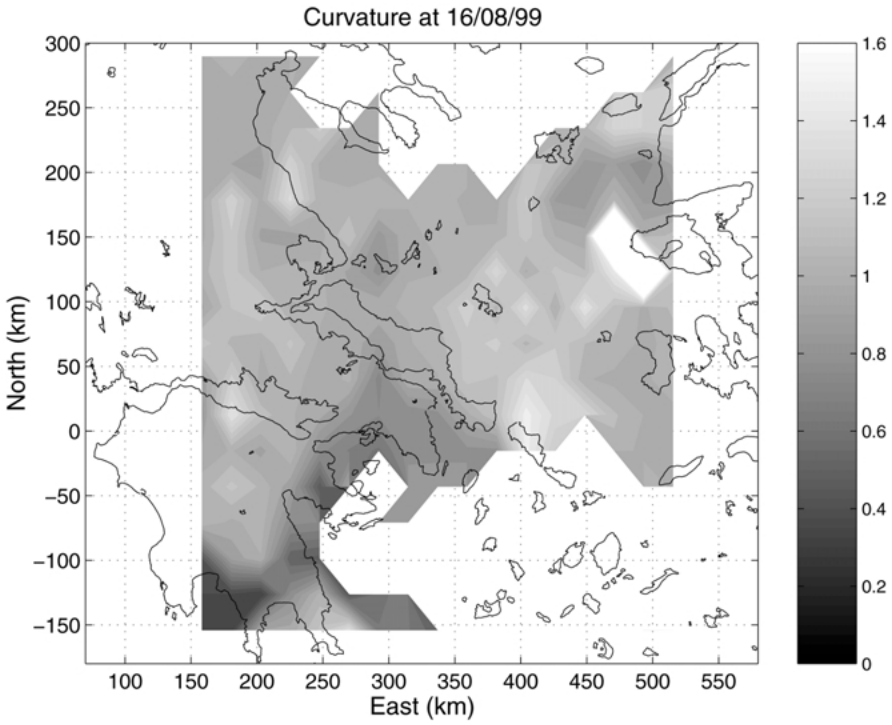

Figure 5. The distribution of curvature computed with data spanning the period 15/6–16/8/1999 (ending just prior to the Izmit event).

from residual stress redistribution and internal deformation in the failed crustal volume, but in the grand scheme of SOC systems external influences are also to be expected. The proportion of internal and external contributions cannot be separated and at any rate, the *short-term* effect of aftershock sequences on regional scaling is neither understood, nor rateable. In consequence, and for the sake of comparison we have produced two curvature maps. The first was compiled using the entire catalogue above the magnitude of completeness, i.e. including the aftershocks (Figure 8a). Then, we applied the Reasenberg (1985) algorithm to identify and remove aftershocks, limiting the search to within 40 km around the epicentre, and re-compiled the curvature map using the declustered catalogue (Figure 8b). Although the two analyses result in somewhat different curvature distributions at the area around Athens, any evidence of power law behaviour has disappeared, meaning that seismic release has reverted to more or less linear (quasi-constant) rates.

The above analysis indicates that prior to 17/8/99, seismic release rates either were quasi-constant, or any power law dependence was subtle and could not be clearly detected. The apparent onset of precipitous power-law behaviour began immediately after the 17/8/99 M7.4 Izmit earthquake (Turkey) and culminated with the Athens event, disappearing soon afterward. They could be clearly ob-

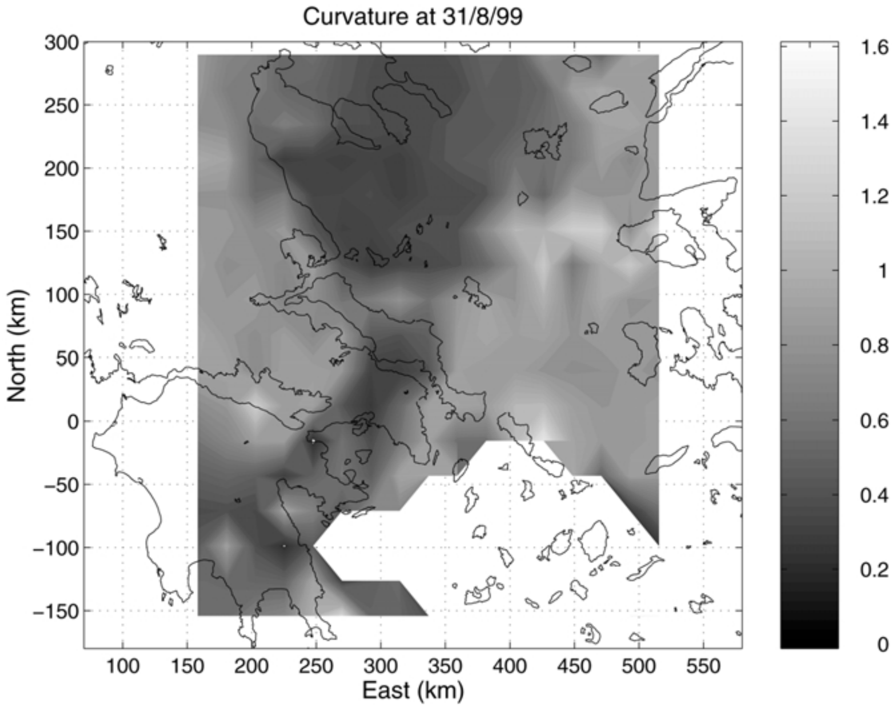

Figure 6. The distribution of curvature computed with data spanning the period 25/6–31/8/1999.

served over the entire North Aegean, the Sporadhes, Euboea, Attica and through the Central and SW Peloponnese. In our view this comprises ample evidence of a causal relationship between the Izmit and Athens events, also involving stress-stress correlation and fault interaction over a broad area. In concluding this section, we also note that evidence of dynamically triggered regional seismicity following the Izmit earthquake was independently presented by Brodsky *et al.* (2000) and Papadopoulos (2000).

5. Discussion

A large earthquake inflicts in the surrounding areas stress changes, which can be static and transient. Static (Coulomb) stresses appear to be effective in the near field of the faults. For instance, more aftershocks occur at regions where static stresses were elevated by a main shock and vice versa (e.g. Reasenberg and Simpson, 1992; Toda *et al.*, 1998; Harris and Simpson, 1998). In addition, there is strong indication that large earthquakes may advance (retard) the time of other large earthquakes by increasing (decreasing) the static stress load in neighbouring areas (see Harris, 1998 for a review). Such examples may be the sequence of 20th century large earthquakes along the North Anatolian fault (Stein *et al.*, 1997), or

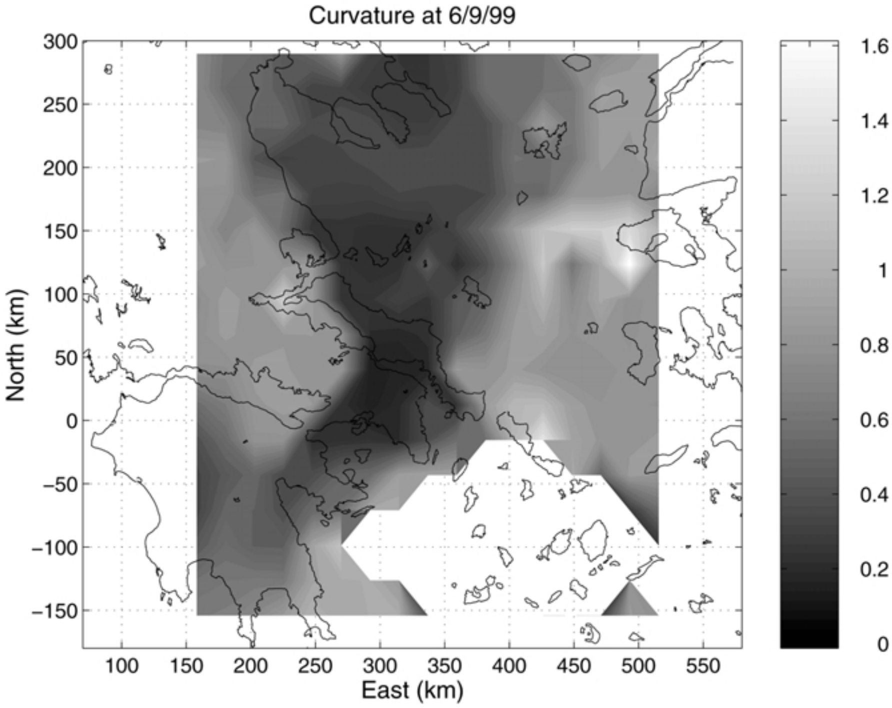

Figure 7a. The distribution of curvature computed with data spanning the period 25/6–6/9/1999, (just prior to the Athens event).

the inferred relationship between earthquakes at and around segments of the North Anatolian fault in the Sea of Marmara region and the North Aegean (Nalbant *et al.*, 1998). It appears however, that all documented cases of earthquake interaction and triggering by static stress changes are either short range, or have intermediate-long time constants (years to decades). Coulomb failure cannot explain observed cases of long range, short term triggered seismicity, occurring at areas where static stress changes are negligible (for instance, Gomberg *et al.*, 1998, and references therein). Remote interactions may possibly be explained with transient effects on the basis of the dynamic rate-and-state friction model, (Harris, 1998; Gomberg *et al.*, 1998), which predicts that transient stress perturbations can have amplitudes orders of magnitude larger than static changes (an illuminating example can be found in Table 3 of Gomberg *et al.*, 1998 and pertinent discussion). According to this model, during the long interval of self-driven accelerating slip, a fault's sliding speed depends on the initial stress and state (fault conditions). After an external stress step is applied due to unstable sliding in another fault, the slip speed depends on the new stress and state. An earthquake (frictional instability) nucleates when the fault strength decreases more rapidly with slip than the elastic stiffness of the surrounding rocks (Gomberg *et al.*, 1998). Thus, a stress increase, (positive step) will advance the time of failure by an amount depending on its magnitude

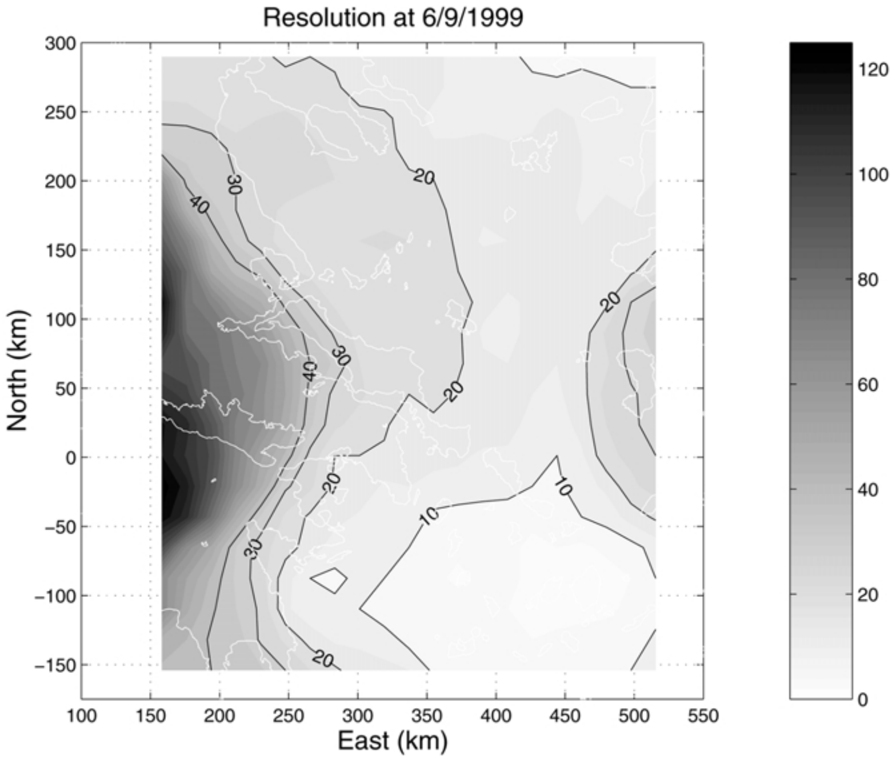

Figure 7b. The *resolution* at 6/9/1999 (number of earthquakes used to calculate the curvature at each grid point).

and fault state, triggering an earthquake if the conditions are right. According to Gomberg *et al.* (1997), near (tens of kilometres) small/moderate earthquakes and remote (thousands of kilometres) earthquakes with magnitudes 2 to 3 units larger may be equally effective at triggering seismicity. Notably, the rate-and-state model is the only able to account for remote interactions (e.g. Harris, 1998) and also predicts that the duration of the increased seismicity rates cannot exceed that of the transient load, consistently with the finite duration of accelerated seismicity observed between the Izmit and Athens events.

However, while dynamic rate-and-state friction may provide a viable explanation of the observed acceleration of seismicity rates and the triggering of the Athens event, it cannot account for its power-law form over such a broad area. The answer may lie in the self-organisation of the regional fault system. A fundamental property of critical systems is their finite hierarchy. As Huang et al. (1998) ascertain, any given level of the hierarchical rupture is equivalent to a critical point for the lower levels. In a CP system without a single largest fault element to rupture, the long range correlation of the stress field may either allow the final event to connect neighbouring faults (as for instance in the case of the 1992 earthquake of Landers, California), or, may simply remove the system from criticality with a number of

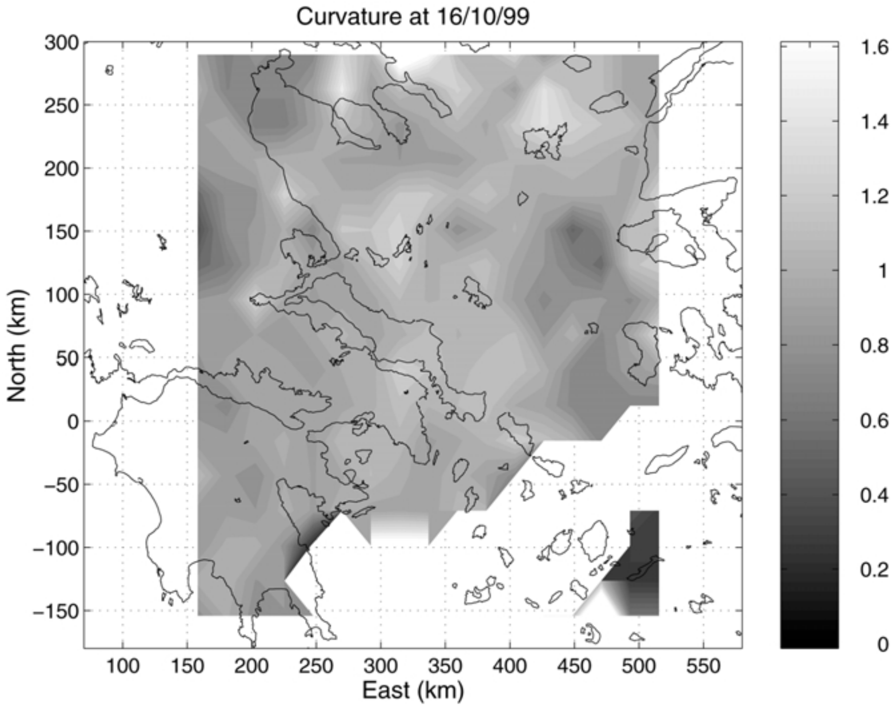

Figure 8a. The distribution of curvature for the period 1/8–16/10/1999, (including the inter-event acceleration period and the post-Athens earthquake days).

small(er) earthquakes. Thus, the transient stress load of the Izmit event may have perturbed the regional fault system, triggering multiple distributed failures through the North Aegean and Continental Greece (we assume that very many more faults were destabilised, than the number corresponding to the $M_L > 2.8$ detectable by the NOA network). The ensuing period of accelerated seismicity can effectively be understood as a re-organisation process, whereby the stress redistribution produced by these failures triggered more faults, until the system reached a new steady state. The many-fault interactions may account for the power-law behaviour of the accelerating seismic release. At Athens, premature failure may have been triggered in a fault nearing its load bearing capacity and supposing that this was the highest level of the perturbed system capable of rupturing at this given time, it may have served as the focal point for the self-(re)organisation process.

In conclusion, the Athens earthquake may have been hastened by long range interaction with the Izmit event. Thus, an important corollary of this work is that following a large earthquake, it is probably wise to speculate for its offspring in its broader region of interaction, as it may be preceded by characteristic changes of seismic release rates. In the reported case of the 7/9/99 $M = 5.9$ Athens earthquake, definite, power-law acceleration of seismic energy release was observed, which could be modelled to yield fair estimates of the critical time, epicentral area and

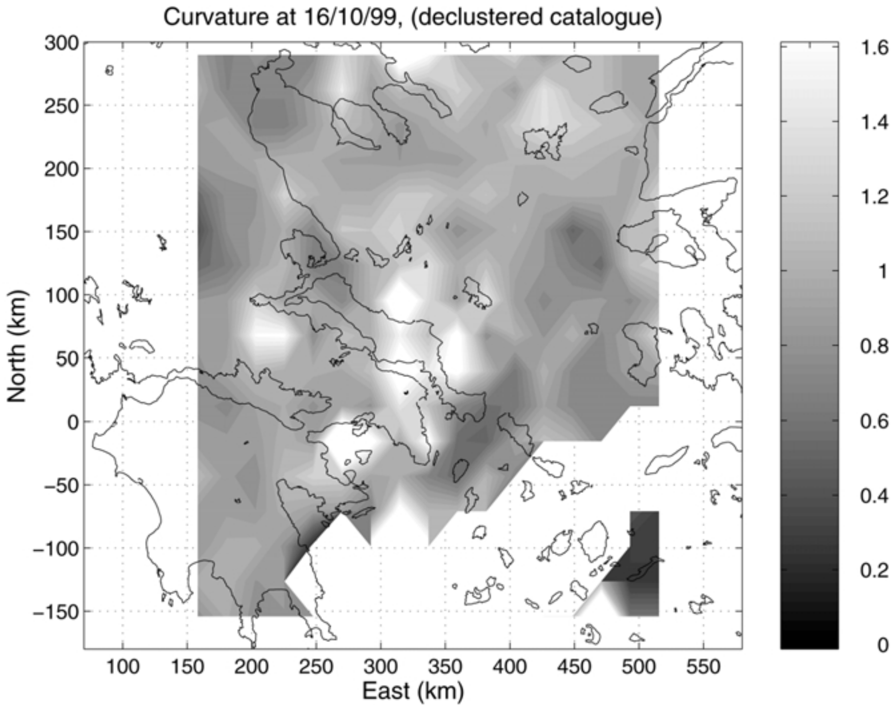

Figure 8b. The distribution of curvature for the period 1/8–16/10/1999, computed from the declustered NOA catalogue (see the text for details).

magnitude of the event. At any rate, the complexity of seismogenetic systems is apparent, as also is the amount of research necessary before we begin to understand them.

References

Allègre, C. J. and LeMouël, J. L.: 1994, Introduction of scaling techniques in brittle failure of rocks, *Phys. Earth Planet Inter.* **87**, 85–93.

Bak, P. and Tang, C.: 1989, Earthquakes as a self-organized critical phenomenon, *J. Geophys. Res.* **94**, 15635–15637.

Bowman, D. D., Ouillon, G., Sammis, C. G., Sornette, A., and Sornette, D.: 1998, An observational test of the critical earthquake concept, *J. Geophys. Res.* **103**, 24359–24372.

Brehm, D. J. and Braile, L. W.: 1998, Application of the time-to-failure method for intermediate term prediction in the New Madrid Seismic Zone, *Bul. Seism. Soc. Am.* **88**, 864–580.

Brehm, D. J. and Braile, L. W.: 1999, Intermediate term earthquake prediction using the modified time-to-failure method in Southern California, *Bul. Seism. Soc. Am.* **89**, 275–293.

Brodsky, E. E., Karakostas, V., and Kanamori, H.: 2000, A new observation of dynamically triggered regional seismicity: Earthquakes in Greece following the August: 1999 Izmit, Turkey earthquake, *Geophys. Res. Lett.* **27**, 2741–2744.

Bufe, C. G. and Varnes, D. J.: 1993, Predictive modelling of the seismic cycle of the greater San Francisco Bay region, *J. Geophys. Res.* **98**, 9871–9883.

Bufe, C. G., Nishenko, S. P., and Varnes, D. J.: 1994, Seismicity trends and potential for large earthquakes in the Alaska-Aleutian region, *Pure Appl. Geophys.* **142**, 83–99.

Czechowski, Z.: 1991, A kinetic model of crack fusion, *Geophys. J. Int.* **104**, 419–422.

Dieterich, J. A.: 1994, A constitutive law for earthquake production and its application to earthquake clustering, *J. Geophys. Res.* **99**, 2601–2618.

Galanopoulos, A. G.: 1956, I seismiki epikindynotis ton Athinon (The earthquake hazard of Athens), *Praktika, Acad. Athens* **31**, 464–472.

Gomberg, J. and Davis, S.: 1996, Stress/strain changes and triggered seismicity at The Geysers, California, *J. Geophys. Res.* **101**, 733–750.

Gomberg, J., Blanpied, M. L., and Beeler, N. M.: 1997, Transient triggering of near and distant earthquakes, *Bull. Seismol. Soc. Am.* **87**, 294–309.

Gomberg, J., Beeler, N. M., Blanpied, M. L., and Bodin, P.: 1998, Earthquake triggering by transient and stativ deformations, *J. Geophys. Res.* **103**, 24411–24426.

Habermann, R., E.: 1983, Teleseismic detection in the Aleutian Island Arc, *J. Geophys. Res.* **88**, 5056–5064.

Harris, R. A.: 1998, Introduction to special section: Stress triggers, stress shadows and implications for seismic hazard, *J. Geophys. Res.* **103**, 24347–24358.

Harris, R. A. and Simpson, R. W.: 1998, Suppression of large earthquakes by stress shadows: a comparison of Coulomb and rate-and-state failure, *J. Geophys. Res.* **103**, 24439–24452.

Herrmann, H. J. and S. Roux (eds.): 1990, *Statistical Models for the Fracture of Disordered Media*, 353 pp., Elsevier, Amsterdam.

Hill, D. P., Johnston, M. J. S., and Langbein, J. O.: 1995, Response of Long Valley caldera to the M_w = 7.3 Landers, California, earthquake, *J. Geophys. Res.* **100**, 12985–13005.

Huang, J. and Turcotte, D. L.: 1990, Are earthquakes an example of deterministic chaos?, *Geophys. Res. Lett.* **17**, 223–226.

Huang, Y., Saleur, H., Sammis, C. G., and Sornette, D.: 1998, Precursors, aftershocks, criticality and self-organized criticality, *Europhys. Lett.* **41**, 43–48.

Jones, L. and Haukson, E.: 1997, The seismic cycle in southern California: Precursor or response?, *Geophys. Res. Lett.* **24**, 469–472.

Kanamori, H.. and Anderson, D. L.: 1975,. Theoretical basis of some empirical relations in seismology, *Bull. Seismol. Soc. Am.* **65**, 1073–1095.

Keilis-Borok, V.: 1990, The lithosphere of the Earth as a large nonlinear system, in G. D. Garland and J. R. Apel (eds), *Quo Vadimus: Geophysics for the Next Generation, Geophys. Monogr. Ser.*, Vol. 60, Washington, D.C.: AGU, pp. 81–84.

Morein G., Turcotte, D. L., and Gabrielov, A.: 1997, On the statistical mechanics of distributed seismicity, *Geophys. J. Int.* **131**, 552–558.

Main, I.: 1997, Long odds on prediction, *Nature* **385**, 19–20.

Nalbant, S. S., Hubert, A., and King G. C. P.: 1998, Stress coupling between earthquakes in northwest Turkey and the north Aegean Sea, *J. Geophys. Res.* **103**, 24469–24486.

Papadopoulos, G. A.: 1997, Earthquake synchronisation as a potential tool for predicting earthquakes, *Annales Geophys.* **15**(1), C49 (abstract, XXII EGS Gen. Ass., Vienna: 1997).

Papadopoulos, G. A. and Drakatos, G.: 1998, A time clustering of strong earthquakes in Greece during October–November 1997, *Annales Geophys.* **16**(4), C1190 (abstract, XXIII EGS Gen. Ass., Nice: 1998).

Papadopoulos, G. A.: 2000, Earthquake triggering in Greece and the case of the 7 September 1999 Athens, Greece earthquake, in Görur and G. A. Papadopoulos (eds), *Integration of Earth Sciences Research on the 1999 Turkish and Greek earthquakes and needs for future co-operative research*, NATO Advanced Research Workshop, Kluwer, in preparation.

Reasenberg, P. A.: 1985, Second-order moment of Central California Seismicity: 1969–1982, *J. Geophys. Res.* **90**, 5479–5495.

Reasenberg, P. A. and Simpson, R. W.: 1992, Response of regional seismicity to the static stress change produced by the Loma Prieta earthquake, *Science* **255**, 1687–1690.

Rundle, J. B, Klein, W., Gross, S., and Turcotte, D. L.: 1995, Boltzmann fluctuations in numerical simulations of non-equilibrium lattice threshold systems, *Phys. Rev. Lett.* **75**, 1658–1661.

Saleur, H., Sammis, C. G., and Sornctte, D.: 1996a, Renormalization group theory of earthquakes, *Nonlinear Processes in Geophysics* **3**, 102–109.

Saleur, H., Sammis, C. G., and Sornette, D.: 1996b, Discrete scale invariance, complex fractal dimensions, and log-periodic fluctuations in seismicity, *J. Geophys. Res.* **101**, 17,661–17,677.

Sammis. C. G., Sornette, D., and Saleur, H.: 1996, Complexity and earthquake forecasting, in J. B. Rundle. W. Klein, and D. L. Turcotte (eds), *Reduction and Predictability of Natural Disasters, SFI Studies in the Sciences of Complexity*, Vol. XXV, Reading, Mass: Addison-Wesley, pp. 143–156.

Scholtz C. H.: 1991, Earthquakes and Faulting: self-organized critical phenomena with characteristic dimension: in T. Riste and D. Sherrington (eds), *Spontaneous formation of space-time structures and criticality*, Norwell, Mass.: Kluwer Academic Publishers, pp. 41–56.

Senatorski, P.: 1997, Interactive dynamics of faults, *Tectonophysics* **277**, 199–207.

Sornette D.: 1991, Self-organized criticality in plate tectonics, in T. Riste and D. Sherrington (eds), *Spontaneous formation of space-time structures and criticality*, Norwell, Mass.: Kluwer Academic Publishers, pp. 57–106.

Sornette, A. and Sornette, D.: 1989, Self-organized criticality and earthquakes, *Europhys. Lett.* **9**,197.

Sornette, D., and Sammis, C. G.: 1995,.Complex critical exponents from renormalization group theory of earthquakes: Implications for earthquake predictions, *J. Phys. 1* **5**, 607–619.

Sornette. D., Vanneste, C., and Knopoff, L.: 1992, Statistical model of earthquake foreshocks, *Phys.Rev. A* **45**, 8351–8357.

Stein, R. S., Barka, A. A., and Dieterich, J. H.: 1997, Progressive failure on the North Anatolian fault since 1939 by earthquake stress triggering, *Geophys. J. Int.* **128**, 594–604.

Toda, S. R., Stein, R. S., Reasenberg, P. A., Dieterich, J. H., and Yoshida, A.: 1998, Stress transferred by the 1995 M_w = 6.9 Kobe, Japan shock: Effect on aftershocks and future earthquake probabilities, *J. Geophys. Res.* **103**, 24543–24566.

Vanneste, C. and Sornette, D.: 1992, Dynamics of rupture in thermal fuse models, *J. Phys. I Fr.* **2**, 1621–1644.

Varnes, D. J.: 1989, Predicting earthquakes by analysing accelerating precursory seismic activity, *PAGEOPH* **130**, 661–686.

Voight, B.: 1989, A relationship to describe rate-dependent material failure, *Science* **243**, 200–203.