Seismic regime of Southern California in relation to the crustal rigidity variations

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Abstract. The Landers (southern California) destructive ($M = 7.3$) earthquake of June 28, 1992 had epicentral coordinates of (34.2°N, 116.44°W) and was a major seismic event of the period from 1985 through 2002. It was examined in many papers, mostly published in a special issue of *Bulletin of Seismological Society of America* (BSSA) of June 1994 (vol. 84, no. 3). The majority of the works were devoted to phenomena related to the earthquake itself and its effects. The present paper presents the results derived from the analysis of rigidity variations at the preparation stage of the main shock. This analysis was based on a method developed by the authors for the monitoring of rigidity variations in the seismically active layer of the crust. The method reduces to the possibility of discriminating between the brittle and viscous failure modes using P wave first arrival patterns.

Introduction

As is generally acknowledged, the high seismic activity of the crust in California is due to the interaction between the Pacific and North American plates. The most prominent structure of the tectonic contact is the San Andreas fault. Right-lateral displacements of the fault walls on certain segments reach a few centimeters per year. It is widely acknowledged that tectonic rhymes of the fault life and seismic activation are interrelated. This idea is based on the hypothesis of *Reid* [1910] known as the theory of elastic rebound. The slowdown of the fault wall motion on certain fault segments leads to the accumulation of elastic strain energy released in the form of an earthquake; the latter restores the overall tectonic situation on the fault. However, the seismotectonic setting at the plate contact is not so simple. The point is that the San Andreas fault does not trend strictly linearly. Thus, in the north, its strike changes from northwest- to westward (at a nearly right angle), and in southern California the fault experiences a bend extending for about 300 km, changing its strike from NW-SE to WNW-ESE. Assuming that vector characteristics of the regional field of tectonic stresses are stationary, strike changes should lead to variations in the relative velocity of fault walls and appearance of extension and compression zones stable in time. Branch faults are very important for the determination of the seismic regime of a region.

A seismic network unique in density, sensitivity and data processing efficiency has been deployed on the territory of the state. A wide energy range of earthquakes recorded (the lower magnitude threshold is less than unity) and a high seismic activity throughout this range are beneficial to the development and testing of new methods of seismic hazard prediction.

This work presents results obtained with a new method of monitoring spatial-temporal rigidity variations in seismically active rock masses of the crust in southern California, where the catastrophic ($M = 7.3$) Landers earthquake occurred on June 28, 1992; the earthquake rejuvenated existing surface faults and formed new ones over more than 60 km. It was preceded by a strong ($M = 6.1$) foreshock (Joshua Tree earthquake) and was followed by an intense series of aftershocks (the $M_{\max} = 6.4$ Big Bear earthquake) that lasted about four years. As mentioned above, the main data on these earthquakes and their consequences have been presented in the special BSSA issue. An earthquake with $M = 7.1$ occurred on October 16, 1999 40 km north of the epicenter of the 1992 event.

The rigidity monitoring method was described in [*Lykov and Mostryukov,* 1996; *Lykov et al.,* 2001] and is based on the following facts.

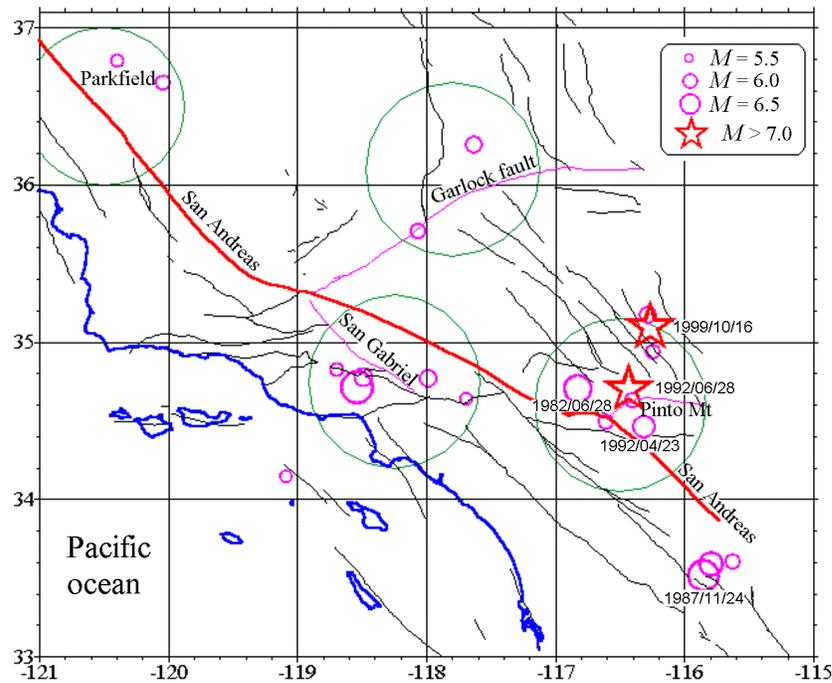(1) Both brittle and viscous types of failure giving rise

Figure 1. Schematic map showing then faults, epicenters of southern California strong earthquakes and areas studied.

to radiation of seismic waves can be observed, with various degrees of probability, in rock masses due to their heterogeneity, regardless of their stress state.

(2) The amount of brittle failure prevails at the stage of elastic deformation, and the viscous failure prevails at the plastic flow stage.

(3) The brittle and viscous types of failure differ in relative velocity of crack edges. This velocity is highest in pull-apart cracks (a variant of the brittle failure).

(4) The relative velocity of crack edges controls the steepness of the leading P wave front, as well as uncertainties in the sign and time determination of first arrivals.

We proposed a new technique allowing the identification of the fracture type. Based on general principles, a possible indicator of the fracture type can be the ratio of the number of stations providing the first arrival sing to the general number of stations that recorded a given earthquake. This ratio that we called the earthquake rating (RG) ranges from 0 to 1 and does not depend on magnitude within the limits determined by the sensitivity and surface density of a network. The following main condition should be met in relation to this problem: an earthquake should be recorded by no less than n stations (in our case $n \geq 10$) around the epicenter. The upper magnitude limit is determined by the requirement that the earthquake recording area should not exceed the seismic network area. In southern California, the magnitude range in question is $0.5 \div 4.0$.

In order to characterize the fracturing process in a seismically active volume, we used in our study the average RG values for earthquakes that occurred over a fixed time interval. The sizes of the standard volume and the time interval length determine the resolution level of the analysis and depend on the seismic regime of the territory studied.

Under ideal conditions, when background earthquakes are uniformly distributed throughout a certain volume of a seismically active rock mass, the average RG parameter provides evidence for the prevailing fracture type and thereby for the integral rigidity of rocks in a given time interval. The scanning technique allows one to perform the following procedures.

(1) Analysis of the time variation RG = $f(t)$ in a fixed volume.

(2) Determination of the surface (in the coordinates φ, λ) RG distribution for a fixed time interval and thickness of the seismically active layer.

(3) Scanning in depth and time of a fixed area $(\Delta\varphi, \Delta\lambda)$.

Results of the Analysis

The earthquakes of 1992 ($M = 7.3$) and 1999 ($M = 7.1$) occurred north of the San Andreas fault (Figure 1), where the background seismicity does not exceed 10 earthquakes per a $2° \times 2°$ and the average background energy level is estimated, on a magnitude scale, as $\sim 2.0 \pm 1.0$. The distance between epicenters of the strongest earthquakes is about 45 km. The information on the background seismicity required for the calculation of the RG parameter is available from seismic bulletins of Southern California Earthquake Data Center stations (http://www.scecdc.scec.org/). In order to gain constraints on the pre- and postseismic stages
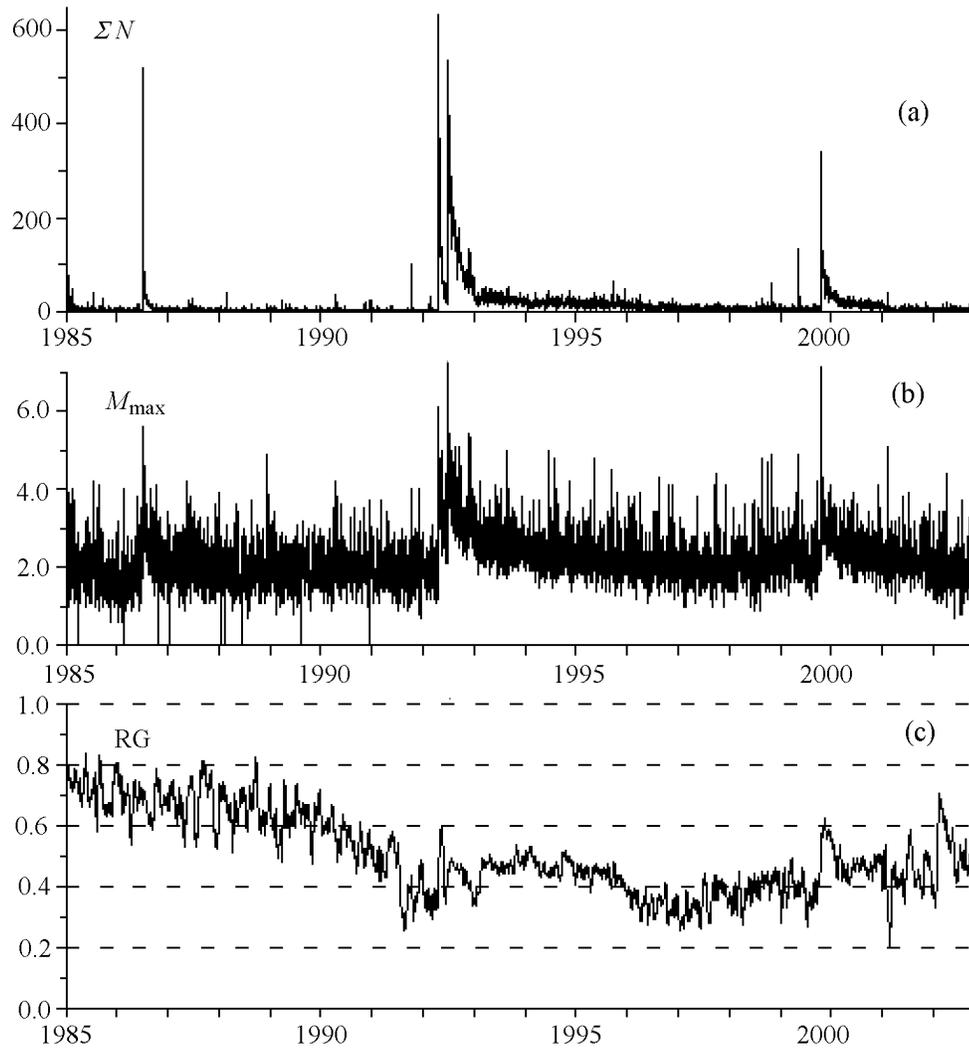
Figure 2. Variations in the parameters of the seismic regime in the Landers area in the period from 1985 through 2002: number of earthquakes pr day (upper plot), maximum magnitude per day (middle plot) and average rating (lower plot).

of earthquakes, we analyzed the data series from January 1, 1985 to April 22, 2002.

Systematic monitoring of the situation in northern and southern California is performed on the basis of current data of seismic networks. The results of RG estimation in several testing areas are regularly displayed on the site http://borok.adm.yar.ru/russian/l_512/index.html in the form of the plots RG = f(t) and maps showing the values of RG and its gradients.

Particular attention in this work was given to epicentral zones of strong earthquakes in southern California mentioned above. Figure 2 plots the 15-day-averaged parameter RG characterizing the crust throughout its thickness (~20 km) in the (33.7°–34.7°N, 115.94°–116.94°W) rectangle. The rating values are seen to systematically decrease since the late 1989. This decrease is particularly pronounced since the mid-1991. The earthquake of April 23,

1992 ($M = 6.1$) with the epicentral coordinates (33.96°N, 116.3°W) occurred 30 km southeast of the main shock against the stable background minimum of RG values (Figure 3). The rating drastically rose after the earthquake. Higher RG values were observed for a month after which decreased once more. The RG values reached their minimum two weeks before the main earthquake that occurred on June 28, 1992 and was accompanied by an RG increase. Although the first of this pair of earthquakes had a smaller magnitude, it was characterized by a more intense aftershock sequence. On the other hand, judging from the parameters of the seismic regime and rating behavior, the aftershock sequence of the main earthquake continued until 1996, when the rating started to systematically decrease (Figure 2).

A nearly linear increase in the rating in 1999 was disturbed by relatively short sign-alternating variations in its values. A somewhat poorly expressed half-year decrease was
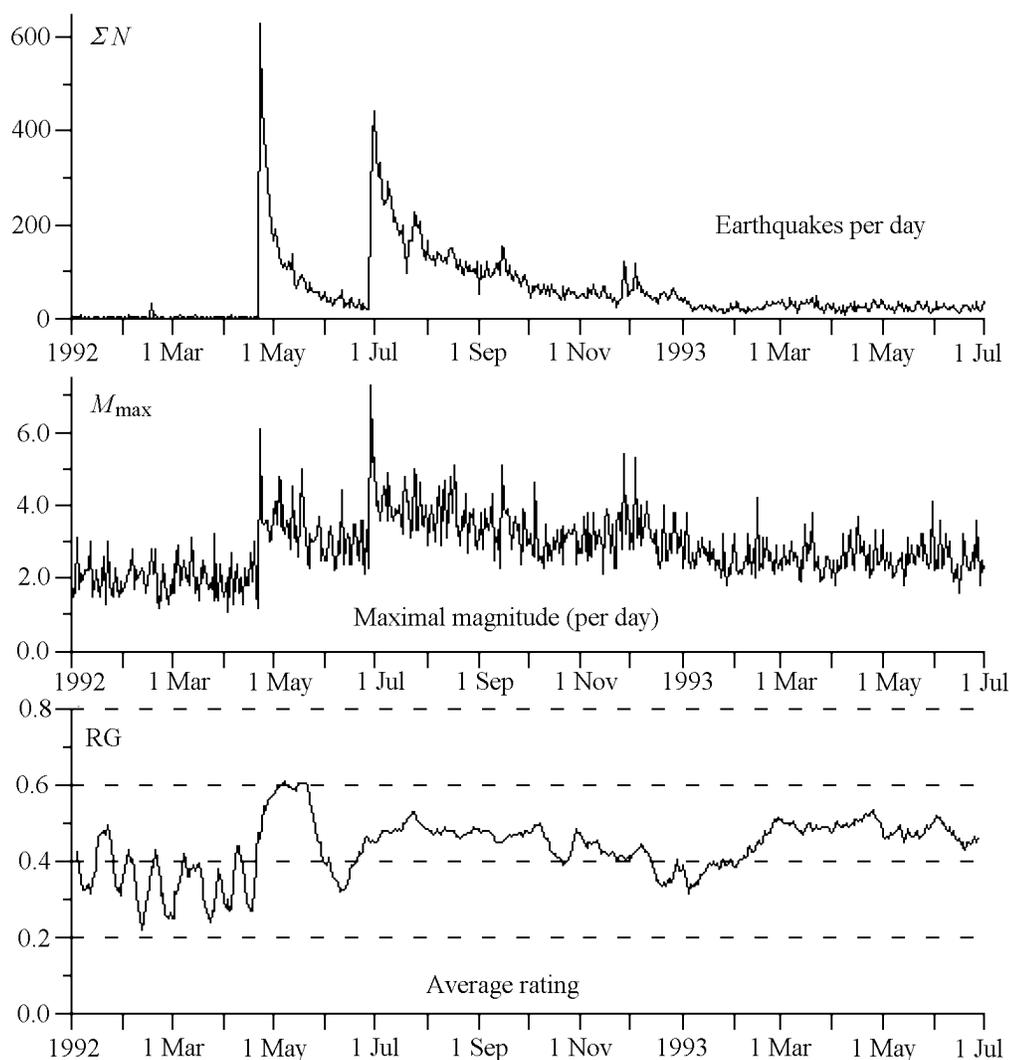
Figure 3. Variations in the parameters of the seismic regime in the Landers area in the period from January 1, 1992 to July 1, 1993: number of earthquakes per day (upper plot), maximum magnitude per day (middle plot) and average rating (lower plot).

followed by a rapid rise when the second earthquake with $M = 7.1$ occurred on October 16. This event did not disturb the linear increasing trend of the rating. In order to constrain the territory where the effect of a rating decrease preceding to an earthquake is observed, the dependences $RG = f(t)$ in three areas are plotted in Figure 4. These areas are located along the active fault system parallel to the San Andreas fault where the earthquakes of 1992 and 1999 occurred. As seen from this figure, the trend component of the rating versus time dependence and variations preceding earthquakes are similar to those observed at the epicenters of both earthquakes (Figure 4a). The centers of the areas are separated in latitude by 1.5°. The areas do not overlap, which guarantees the independence of weak earthquake samples that were used for the RG determination. Figure 4c characterizes the Long Valley volcanic caldera. The center of this area lies at a distance of 2° northward from the area

of Figure 4b. As can be seen, the events in the southern area had no effect on the RG trend. Neither any sharp decrease in the rating is observed in 1991–1992. Therefore, precursory anomalies preceding the earthquakes of 1992 and 1999 encompassed a limited area. Similar check calculations were performed for the San Andreas fault zone. Thus, whereas the situation in the San Gabriel fault zone of this system is completely analogous to the Landers and Garlok-fault areas, the time behavior of the rating in the Parkfield area (1.5° north of San Gabriel) and Loma Prieta (1° north of Parkfield) has no signatures of active processes in the southern California region. The size of the zone of precursory anomalies is roughly estimated at 300 km, which is consistent with the results of *Dobrovolsky et al.* [1979], who estimated the sizes of preparation areas of earthquakes of various magnitudes based on the development of long-term precursors. An abrupt decrease in the rating in the period from the
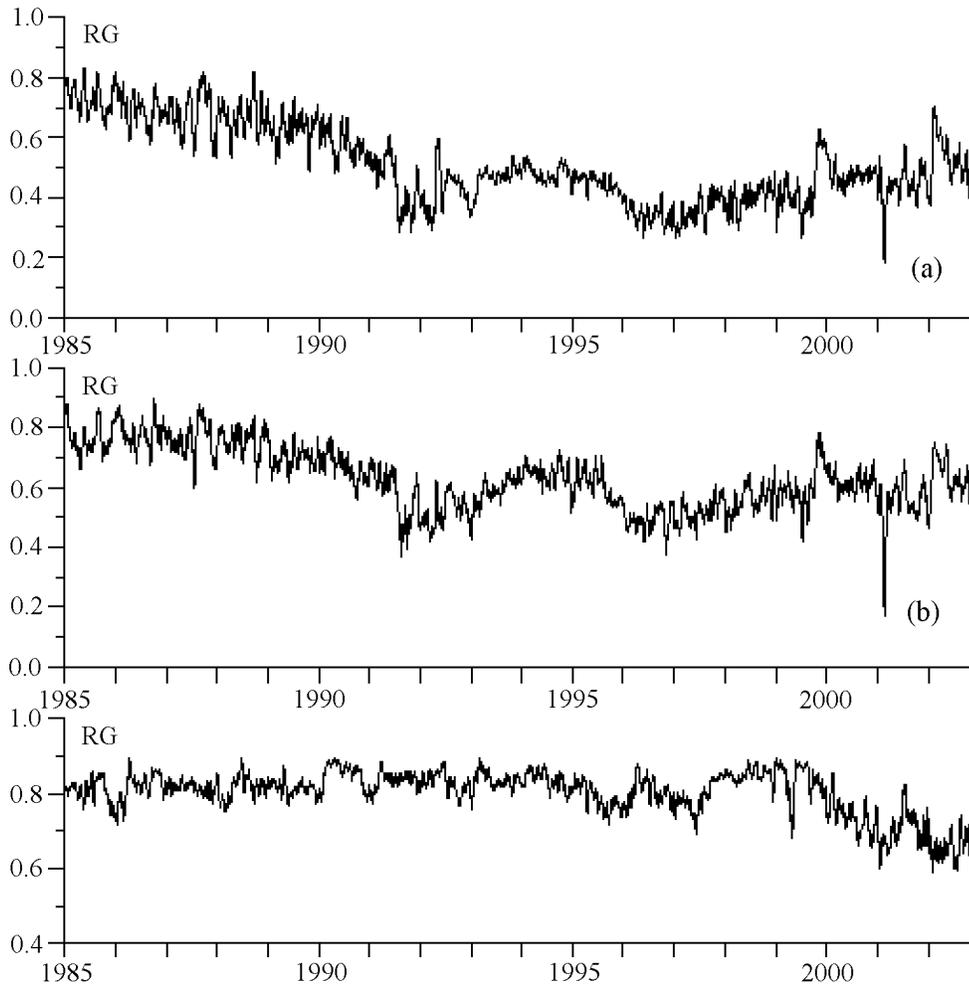
Figure 4. Variations in the rating of the California earthquakes: A, Landers (116.5°W, 34.1°N); B, Garlok fault (117.8°W, 35.6°N); C, Long Valley (118.7°W, 37.6°N).

mid-1991 through April 1992 that can apparently be classified as a medium-term precursor was most pronounced between the Garlok and Pinto Mountain faults and coincided in plan view with Central Mojave Domain blocks [*Unruh et al.,* 1994]. The area of lower rating values was bounded to the southeast by the Pinto Mountain fault, south of which lies the zone of higher rating values observed immediately before the earthquake of April 23, 1992, which occurred precisely in this zone (Figure 5). The main, June 28, 1992 earthquake occurred north of the Pinto Mountain fault. The relationship between these two earthquakes illustrated in Figure 2 allows one to classify the earlier earthquake as a foreshock.

A specific feature of the earthquakes of 1992 and 1999 is the shallowness of their sources (1 and 0 km, respectively). We propose the following model describing the processes that contributed to the occurrence of these earthquakes at various depths of the seismically active crustal layer of no more than 25 km in thickness in their epicentral areas. Most convenient for this purpose is the scanning diagram RG = f(h, t) obtained in the epicentral areas of the earthquakes of 1992 and 1999 (Figure 6). The area size (2°×2°)

is chosen under condition that the number of background earthquakes in a 0.5-km thick crustal layer should be representative for the analysis. The scanning steps in depth and time were 0.1 km and 0.05 yr, and the respective widths of averaging windows were 0.5 km and 0.2 yr. The scanning diagram shows that the decrease in the rock rigidity preceding the earthquake of 1992 started with surface layers from 1989 and involved the entire 20 km of the seismically active crustal layer by 1991. Against this background, an abrupt rating decrease associated with the earthquakes of 1991–1992 penetrated nearly synchronously the entire layer. The Joshua Tree, April 23, 1992, $M = 6.1$ earthquake occurred to the south of the main shock epicenter in the zone of relatively higher RG values and dramatically increased the general rating level. Therefore, the main earthquake occurred against the background of increasing rigidity of rocks of the seismically active layer, and higher RG values persisted until the end of 1995, i.e. until the end of the aftershock activity. The rating level synchronously decreased at all depths in the late 1995-early 1996, after which it started to recover. Starting at great depths, the recovery process gradually involved
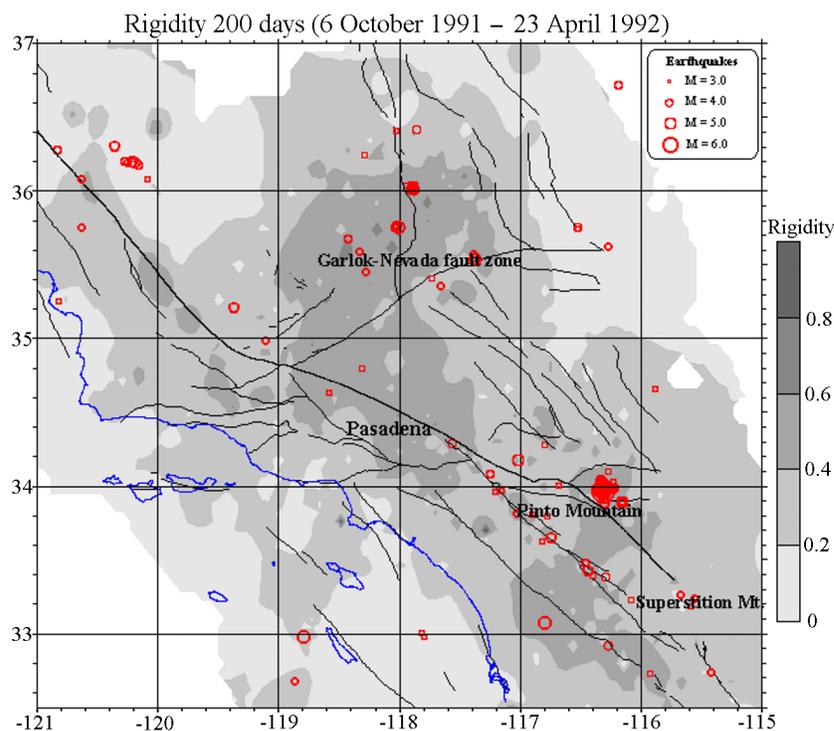
Figure 5. Spatial distribution of the rating over a 200-day interval before the earthquake of June 28, 1992.

shallower depths. An abrupt rating decrease observed at all depths took place three months before the $M = 7.1$ earthquake of October 16, 1999. By this time, the rating level was still not recovered in the near-surface layer. The earthquake was accompanied by a rigidity increase at the surface. In this case, one should speak of rupturing points in the source rather than its depth, because the source depth can be assessed from both the thickness of the crustal layer involved in active processes of rigidity variations and the depths of strong aftershocks.

Discussion

Prehistorical processes of the strong earthquakes of 1992 and 1999 had their signatures in the data of the rock rigidity monitoring on the basis of information gained from first P wave arrivals from background earthquakes. As was discovered in [*Lykov and Mostryukov,* 1996; *Lykov et al.,* 2001], a decrease in the rating of background earthquakes preceding a strong event is its precursor and, in our opinion, this is evidence of a decrease in the integral viscosity of a certain volume of the seismically active rock mass. The term "rigidity" is not used here in its purely mechanical meaning because, as was shown by *Savich* [1983], the rigidity of a rock mass differs from its laboratory determinations in rock samples by a few times. A reduction in the amount of brittle fracture in a rock mass indicates that deformation is effected through displacements on active faults, resulting in

the consolidation of the rock mass. A main earthquake occurs precisely against the background of increasing rigidity. This phenomenon, which we discovered for the first time by analyzing data from the Beijing seismic research area, precedes earthquakes in regions with any tectonic structure [*Lykov et al.,* 2001].

When applying the given method of monitoring the rock rigidity, the following circumstances should be taken into account.

(1) The use of the rigidity parameter for the regionalization of a seismically active territory should take the seismic network pattern into account because the earthquake rating is naturally underestimated outside the network area as background earthquake epicenters become farther from stations.

(2) The analysis of time series RG $= f(t)$ in test areas should take into account both the effect of strong earthquakes occurring outside the area on the RG behavior and cases of the tectonic activation of major faults. Thus, in our case, the November 24, 1987 ($M = 6.6$; 33.01°N, 115.85°W) earthquake in the San Andreas fault zone affected the behavior of the dependence RG $= f(t)$ in the epicentral area of the Landers earthquake (Figure 2).

Conclusion

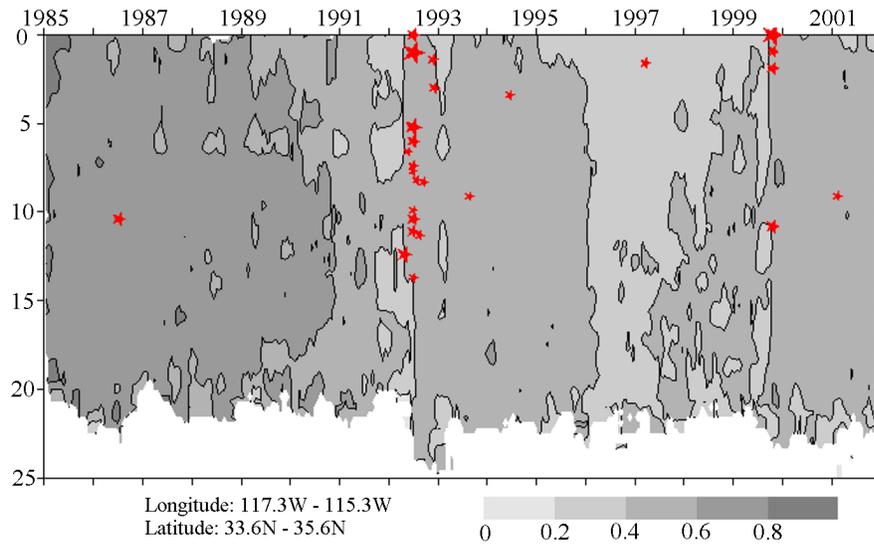The case study of strong earthquakes of 1992 and 1999 in southern California showed that monitoring the rigidity

Figure 6. Scanning diagram RG $= f$(h, t) for a $2° \times 2°$ area centered at (116.3°W, 34.6°N). Stars are $M > 5.0$ earthquakes.

of seismically active rock masses proved to be effective for tracing preseismic processes. The following results have been obtained.

(1) A precursor of an $M = 7.3$ earthquake in the form of a bay-like decrease in the integral rigidity arose in a limited area (about 300 km across).

(2) The rigidity decrease preceding the earthquake of 1992 originated in surface layers of the seismically active sequence and successively, with an acceleration, propagated throughout the entire sequence to a depth of 25 km. This fact appears to be related, in a way, to a very shallow depth of the seismic source (1 km). On the other hand, an abrupt decrease in rigidity observed one year before the earthquake occurred nearly synchronously throughout the depth range; incidentally, the $M \geq 5.0$ aftershock depths reached 15 km. This indicates that the preparation of the earthquake source involved the entire seismically active portion of the crust, and the rupture started to propagate at a depth of 1 km.

(3) The preparation of the 1999 earthquake source started with a rigidity decrease that was synchronous at all depths, after which the rigidity was successively recovered in the upward direction toward the surface. In this case, the instrumentally determined depth of the source was also very small (0 km). Similar to the earthquake of 1992, a rigidity decrease throughout the thickness of the seismically active crust was observed during four months, and the depths of its strong aftershocks exceeded 10 km.

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