

Medium-term prediction of earthquakes from a set of criteria: Principles, methods, and implementation

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Abstract. This paper presents results obtained in 1980–2002 from studies of various aspects of earthquake prediction in relation to the development of a method of medium-term prediction from a set of criteria.

1. Introduction

By their catastrophic consequences, number of victims, material damage, and adverse environmental impact, earthquakes are the most dangerous natural catastrophes. They are not so dangerous by themselves as due to the fact that they occur in populated areas. Their sudden onset aggravates even more their destructive consequences. Damage and casualties are due to not only ground motions proper but also various secondary natural phenomena activated by earthquakes (creep, landslides, rockfalls, snowslides, ground liquefaction, and so on). Secondary anthropogenic effects and aftermath (fires, explosions, releases of radioactive and toxic matter, and epidemics) are also highly dangerous.

Such natural phenomena as earthquakes are inevitable and cannot be prevented, but their destructive effects can and must be mitigated. For this purpose, one should know origins of earthquake occurrence, study processes related to the preparation and realization of earthquakes, and develop methods for their prediction.

Earthquakes account for 13% of total number of natural catastrophes that occurred in the world from 1965 through 1999, occupying the third place [Osipov, 2001]. According to data of the National Earthquake Information Center (NEIC) of the United States, 2000 earthquakes with magnitudes $M_S \geq 7.0$ occurred in the world over the 20th century (from 1900 through 1999), and 65 of these had magnitudes $M_S \geq 8.0$. Casualties associated with the earthquakes of the 20th century amounted to 1.4 million victims. Of this number, 987 thousand casualties (about 32900 victims per

year) account for the last 30 years, when human and economic losses were fixed more accurately. According to data of the Center for Research on the Epidemiology of Disasters (CRED) over 1965–1999, among all natural catastrophes, earthquakes occupy third place by the number of deaths (17% of the total number).

Over 25% of Russia's area belongs to seismically hazardous zones, where seismic tremors of an intensity of 7 and more are likely to arise [Ulomov, 2000; Ulomov and Shumilina, 1999–2000]. This territory includes 3000 large and small cities and settlements, 100 large hydro- and thermoelectric stations, 5 nuclear power plants, and numerous installations of higher ecological hazard. The territory of 103 Russian cities is earthquake prone [Osipov, 2001; Ulomov and Shumilina, 1999–2000]. The above implies the importance and relevance of earthquake prediction as part of a more general task of mitigating the hazard and economic aftermath of natural catastrophes.

Notwithstanding a great number of precursors, none is capable of giving exact information on the time, place and strength of a forthcoming earthquake. Various precursors differ in their behavior in various seismically active regions and yield a large scatter in estimates of the time, place and strength of a future earthquake. Note also that reports on earthquake precursors are generally and it is difficult, if possible at all, to use this evidence for estimating, even retrospectively, their statistical characteristics such as the probabilities of correct prediction and false alarm and the average expectation time of an earthquake after recording of a precursor.

Analysis of long-term data on several geophysical (mostly seismological) precursors showed that the probability of successive prediction using each of the precursors does not exceed 0.5 [Zavyalov, 2002]. A possible way out of this situation is the joint use of several prognostic criteria. Practice of last years showed that, notwithstanding its disadvantages, this approach is justified, at least in relation to medium-term (a few years) prediction.

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Methods applied to the solution of this problem are based on the following principles:

- physical substantiation of the precursors in use, implying constraints on the relation of the precursors to the preparation process gained from theoretical and experimental results;
- estimation of basic statistical characteristics (probability of correct prediction, false alarm probability, and so on) and prognostic efficiency of the precursors used;
- mapping of the precursor values in space and time;
- Bayesian approach to the determination of probabilistic prediction characteristics in the multiple-criteria approach.

2. Brief Review of Earthquake Prediction Algorithms

An earthquake prediction algorithm is understood here as a sequence of operations intended for the identification of characteristic criteria or anomalous variations in rate of geophysical-geological fields and their joint treatment and analysis for the determination of the place, time and strength of a future earthquake.

KORA-3 algorithm. The first stage of scientific research on earthquake prediction mainly consisted in accumulation of empirical data on phenomena preceding the onset of strong earthquake and disappearing after its occurrence. These criteria can be classified as *dynamic*. A few tens of cases of these phenomena had been recorded by the late 1960s–early 1970s. On the other hand, based on a set of geological-morphological and geomorphostructural criteria in seismically active regions, researchers tried to identify areas capable of generating strong earthquakes; these criteria can be classified as *static*. A theoretical prerequisite for the use of data on contemporary topography was the idea that large landforms in seismically active regions were produced by orogenic tectonic processes involving lithospheric depths comparable to hypocentral depths of earthquakes [Gelfand *et al.*, 1976a].

For the first time, this problem was formulated as a problem of predicting places of strong earthquakes by V. I. Keilis-Borok in the early 1970s [Gelfand *et al.*, 1972a; Keilis-Borok and Kossobokov, 1984, 1986; Keilis-Borok *et al.*, 1980; Rantsman, 2001]. It was attacked not only by geophysicists but also by geologists and mathematicians (E. Ya. Rantsman, I. M. Gelfand, Sh. A. Guberman and M. L. Izvekova) [Gelfand *et al.*, 1972a, 1973, 1974a, 1974b, 1976]. Afterward, large contributions to the solution of the problem of recognizing places of strong earthquake occurrence were made in [Bhatia *et al.*, 1992; Caputo *et al.*, 1980; Cisternas *et al.*, 1985; Gelfand *et al.*, 1972b, 1976b; Gorshkov *et al.*, 1979, 1985, 1991, 2001; Gvishiani *et al.*, 1982, 1984, 1988; Healy *et al.*, 1992; Kossobokov and Khokhlov, 1993; Kossobokov and Rotvain, 1977; Kossobokov *et al.*, 1990, 1999; Romashkova and Kossobokov, 2001, 2002; Zhidkov *et al.*, 1975].

Based on the hypothesis according to which epicenters of the majority of strong earthquakes gravitate toward morphostructural nodes, the latter were chosen as objects for recognition. A set of criteria (characteristics of higher and lower seismic activity of nodes) proved rather extensive, amounting to a few tens of parameters. All these parameters were geological and characterized intense and contrasting neotectonic movements and tectonic fragmentation of nodes. Using a special set of criteria, an algorithm known under the name KORA-3 [Bongard, 1967] based on the pattern recognition method was first to be trained, i.e. to select criteria specific to high seismicity nodes (*H*) and to low seismicity nodes (*L*).

Instrumental data on geophysical and geochemical fields, as well as on weak seismicity, were not used, although their suitability for recognition was not denied [Gorshkov *et al.*, 2001]. The authors did not present any physical arguments in favor of one or other criterion [Gelfand *et al.*, 1972a]. The training procedure used samples of *H* and *L* nodes at which earthquakes of predicted magnitudes occurred or, respectively, did not occur in the past.

The second stage consisted in the separation of nodes, at which no strong earthquakes had been previously recorded, into the *H* and *L* categories. The difference (Δ) between the numbers of the *H* and *L* criteria fixed for a given node was used as a parameter based on which the node was attributed to the *H* or *L* type. A threshold value of Δ was selected experimentally. The first results obtained by this technique were published in [Gelfand *et al.*, 1972a].

Over the period from 1972 through 2000, the recognition procedure was applied to more than 1100 morphostructural nodes. In all, 68 strong earthquakes occurred over 30 years in all of the regions studied. 57 (84%) of these events occurred at nodes that were recognized as *H* nodes, 4 (6%) events occurred at *L* nodes, and 7 (10%) events occurred outside the nodes (target missing). Most of the latter cases were confined to lineament zones.

An advantageous compromise between geomorphological and mathematical methods of the recognition of possible earthquake occurrence places that was attained in these studies has given good results. A drawback of this approach is its total disregard for the time aspect; i.e. places were predicted in which strong earthquakes of given magnitudes should sooner or later occur. In other words, the static problem of predicting the place and strength of an earthquake was solved. The question of the occurrence time (the third component of prediction) remained open. However, in our opinion, the recognition problem can also be formulated on the basis of dynamic time-dependent criteria. No theoretical appear to be present in this way.

Another drawback of this approach is inadequate physical substantiation of the relation between recognition criteria and the process of the strong earthquake generation; as mentioned above, we mean the set of principles that are consistent with theoretical and experimental data, providing constraints on the relation of precursors to the preparation process.

An advantage of the prognostic procedure based on the KORA-3 algorithm is its good formal description, admitting an independent verification. We should also note that

KORA-3 was the first algorithm intended for prediction of strong earthquakes and using a set of prognostic characteristics.

Fortran Generalized Portrayal algorithm. The inability to predict the earthquake occurrence time from data on statistical criteria, on the one hand, and a large number of dynamic prognostic criteria (their recording time was related, in a way, to the earthquake occurrence time) accumulated by the early 1970s, on the other hand, induced researchers to seek other ways for solving the pattern recognition problem in order to predict not only the place of a future earthquake but also the time (or time interval) of its possible occurrence.

In the 1970s, a theory and methods based the pattern recognition techniques were developed for the reconstruction of time dependences from empirical data [Vapnik, 1979, 1984; Vapnik and Chervonenkis, 1974]. Unlike the KORA-3 algorithm, this method is statistical, uses the principle of empirical risk minimization and, which is important for the prediction problem, allows one to work with small samples. An algorithm implementing this method was called the Fortran Generalized Portrayal (FGP) algorithm [Vapnik, 1984].

The FGP algorithm, implementing the solution to the pattern recognition problem, is based on the generalized portrayal method of the construction of the hyperplane separating the space of given criteria into one or other situation (in the simplest case, into two classes) [Vapnik, 1979, 1984; Vapnik and Chervonenkis, 1974]. The generalized portrayal means in this case the minimum-modulus directing vector of the separating hyperplane. Since all computer programs implementing the method of generalized portrayal were written in FORTRAN, this is reflected in the method's name.

The FGP method, intended for the recognition of spatial cells and time intervals of possible earthquakes of given energy classes, was applied for the first time to the Caucasus in the early 1990s [Chelidze et al., 1995; Construction ..., 1991]. Initial data for this study were taken from the Caucasus regional catalog of earthquakes. The recognition problem was solved with a set of strong ($K \geq 12.5$) earthquakes (and their groups) and a set of prognostic criteria determined from weak seismicity data processing: the recurrence plot slope γ , density of seismogenic faults K_{sf} , number of earthquakes n , and released seismic energy $E^{2/3}$. All criteria were represented as deviations of their current values from their long-term (background) values and were normalized to the rms error of the determination of the long-term series average. Values of these parameters were calculated in a moving time window $\Delta T = 3$ years long with a step of $\Delta t = 3$ months in rectangular cells 50×50 km and 100×100 km in size with a half-size overlapping. The informativeness of each criterion was checked separately before applying the recognition procedure to the set of criteria.

The recognition problem was formulated in terms of the identification of higher seismic hazard (HSH) time intervals. A six-year interval preceding a strong earthquake was specified for each cell in which such an earthquake occurred. Results of the HSH time interval recognition based on the use of the precursor γ and the joint use of the parameters γ and $E^{2/3}$ showed that correct recognition of the $K \geq 12.5$ earthquake occurrence was attained in 74% of all cases.

The method of the HSH time interval identification based on the FGP algorithm is devoid of the main drawbacks of the method using the KORA-3 algorithm and is effective for predicting not only the place and strength of a future earthquake but also the time (time interval) of its occurrence. Moreover, parameters used as precursors have a clear physical meaning and are related to the preparation process. Unfortunately, the FGP method has not been further developed in relation to earthquake prediction problems. This was largely due to the breakup of the USSR and the loss of scientific contacts in the mid-1990s.

Complex of the California-Nevada (CN), magnitude 8 (M8) and Mendocino scenario (MSc). These algorithms were designed and developed in [Allen et al., 1984, 1986; Keilis-Borok and Kossobokov, 1984, 1986]. These algorithms continue and develop approaches underlying the problem previously formulated for the recognition of possible strong earthquake occurrence places [Gelfand et al., 1972a]. They realize prediction of not only the place but also the time and magnitude of a potential earthquake. The development of the CN, M8 and MSc algorithms was apparently stimulated by the inability of the KORA-3 algorithm to predict the earthquake occurrence time. They use a complex of prognostic criteria determined primarily from data on seismicity.

A final goal of all algorithms is the determination of times of increased probability (TIP) for predicting strong earthquakes with magnitudes $M \geq M_0$ based on analysis of a complex of earthquake flow properties. Values of precursory parameters were calculated from data of a catalog of earthquakes, aftershocks of which were preliminarily removed by the method proposed in [Keilis-Borok et al., 1980]. The CN and M8 algorithms used the same set of formal prognostic criteria determining various properties of the earthquake flow [Sadovsky, 1986]. The physical meaning of the prognostic functions in use and their relation to earthquake preparation processes were not considered by the authors of the algorithms, who were mainly guided by the experience of retrospective statistical analysis of catalogs. The strong earthquake magnitude value M_0 varied from one region to other but was not lower than 6.4.

Retrospective testing of the CN and M8 algorithms in various seismically active regions of the world [Sadovsky, 1986] has shown the following. 23 of 29 strong earthquakes recorded over the testing time occurred in the CN predicted TIP intervals whose overall length amounted to 23% of the total testing time.

Eight of nine strong earthquakes occurred in the M8 predicted TIP intervals. The overall length of the latter amounted to 16% of the total testing time, whereas the overall expectation time accounted for 11% [Sadovsky, 1986].

Notwithstanding good results of its application, the CN algorithm has not been further used. This might be attributed, among other factors, to the fact that the alarm state was declared in this algorithm over large territories (a few hundred thousands of squared kilometers).

Although the same drawback characterizes the M8 algorithm, Healy et al. [1992] and Kossobokov and Khokhlov [1993] reported on the beginning of experimental medium-term prediction of $M \geq 7.5$ earthquakes that would be con-

ducted in a real-time mode with the use of the M8 algorithm. This experiment is still continuing.

Large areas of scanning cells in which TIP intervals are sought induced the authors of the M8 algorithm to develop an auxiliary algorithm reducing the spatial uncertainty of prediction. This algorithm, called the Mendocino scenario (MSc) [Kossobokov *et al.*, 1990], is applied if a TIP period is diagnosed in a testing area of size D with the use of CN or M8. Then, this area is divided into square cells S_{ij} of the linear size $D/16$. Relative seismic quiescence in the TIP period (with respect to previous time) is sought in each of the resulting cells S_{ij} . The authors of the MSc algorithm note that such anomalous quiescence periods cannot be identified in a low-activity part of the alarm territory. Therefore, the MSc algorithm identifies precisely the areas of higher seismic activity, immediately involved in the earthquake preparation process [Romashkova and Kossobokov, 2001].

Data of the MSc algorithm testing in both retrospective approach and real-time mode showed that the MSc application complementing CN and M8 reduces the spatiotemporal volume of alarm by three to four times for earthquakes of $M_0 \geq 8$ and by seven to eight times for earthquakes of $M_0 \geq 7.5$.

Thus, all algorithms considered above are based on the recognition of situation patterns characterized by a set of criteria statistically related to possible generation of strong earthquakes and their preparation process. The algorithm outputs present a list of places, zones and territories in which strong earthquakes of a given magnitude range (KORA-3 algorithm) or a list of time periods during which strong earthquakes of a given amplitude range can occur in a certain seismically active territory (FGP, CN and M8 algorithms).

The algorithms use a complex of prognostic characteristics. However, most of the latter do not have a distinct, physically substantiated relation to the earthquake preparation process. The authors proceeded from intuitive considerations that this or other criterion reflects, in a way, the preparation process. Recognition results obtained by the algorithms confirm these considerations only statistically but do not clarify the physics of the process itself.

A drawback of the KORA-3 algorithm is its inability to predict the occurrence time of a strong earthquake. KORA-3 solves a static problem of predicting the place of a strong earthquake. A significant drawback of the CN and M8 algorithms consists in large sizes of their spatial scanning cells, which is due to sufficiently large samples of earthquakes required for effective operation of the algorithms.

None of the algorithms provides mapped information on the occurrence probability of a strong earthquake at a given point of the seismically active territory under consideration.

3. Earthquake Precursors and Physical Principles Underlying Earthquake Prediction

Earthquake preparation models. The majority of preparation models were developed in the 1950s–1970s. This

was primarily associated, on the one hand, with the interest (increased in the mid-1950s–1960s) to the earthquake prediction problem and its formulation as a fundamental scientific problem [Gamburtsev, 1954, 1982] and, on the other hand, and the discovery of macrofracture precursors of various physical origins observed in both laboratory experiments with samples of model materials and in situ investigations in seismically active regions [Sadovsky and Nersesov, 1978]. A feature common to the vast majority of the earthquake preparation models was their qualitative approach to the description of the process [Zubkov *et al.*, 1980].

Avalanche-unstable fracturing (AUF) model [Myachkin *et al.*, 1975]. Theoretical and experimental prerequisites of its development were principles of solid fracture mechanics based on the kinetic concept of strength [Zhurkov, 1968; Zhurkov *et al.*, 1977] and achievements of observational seismology and other geophysical disciplines.

According to the AUF model, the earthquake preparation process consists of three stages.

Stage 1: quasi-homogeneous cracking. Growth of existing cracks and formation of new ones takes place at this stage due to tectonic stresses. This process promotes the development of initial conditions for the formation of a main rupture but is not accompanied by phenomena of the precursory type.

Stage 2: avalanche-like interaction of cracks. This stage begins when a certain critical density of cracks is attained in the preparation volume: the interaction between the cracks gives rise to cracks of larger sizes and rapid abrupt variations in local stress fields. This process is accompanied by variations in integral characteristics of the medium that are observed in various geophysical fields and are precursors of the nucleating earthquake.

Stage 3: instability stage. Deformations are localized in a narrow region encompassing the zone of the future main shock; several relatively large cracks develop in this region. As a result of accelerating growth of deformations in this region, a stress drop and elastic recovery of physicomaterial properties are observed in the surrounding volume. The preparation zone divides into elastic and inelastic subzones. The narrow zone of unstable deformation is characterized by a higher concentration of fractures, and their set forms the main fault surface. This fault (the earthquake source) is produced through the breakage of bridges between large cracks. Qualitatively, this process develops in a similar way at various scale levels, i.e. the formation of the large main rupture is advanced by similar acts of smaller fracturing. Later, this was confirmed under conditions of laboratory experiments, mining operations, and in situ observations in seismically active regions [Dyagilev, 1998; Malovichko *et al.*, 1998; Zavyalov and Nikitin, 1997a, 1997b]. The scale invariance of the fracture process is an advantage of the AUF model. Apparently, the AUF model served as a cornerstone of the modern ideas representing a rock as a block-fragmented hierarchical self-organizing medium [Sadovsky, 1979; Sadovsky *et al.*, 1982].

Ideas similar to the AUF model were advanced in works by W. Stuart and B. Brady in relation to earthquake preparatory processes. As distinct from the AUF model, the

strain localization zone in the model of *Stuart* [1974] coincides with the fault present and fractured rock in the vicinity of the latter behaves, on a macroscopic scale, as a material with a dropping stress-strain diagram. The strain development region in the model of *Brady* [1974, 1975, 1976] is considered as a soft elastic inclusion with lower effective elastic moduli; the main rupture develops due to stress redistribution such that a favorably oriented major crack growth and other cracks close.

The presence or absence of water in the preparation region is inessential for the AUF, *Stuart* and *Brady* models. On the contrary, in the dilatancy-diffusion (DD) theory of earthquake preparation developed by *Scholz et al.* [1973] and likewise based on the fracturing process, tensile fractures prevail and the presence of water in rocks of the hypocentral region is a factor of basic significance.

According to the DD model, macrostresses gradually increase at the first preparation stage characterized by elastic behavior but neither opening of existing cracks nor formation of new ones occurs. At sufficiently large values of differential stresses (larger than 50% of the failure value), numerous tensile cracks arise and a relative increase in the rock volume (dilatancy) is observed. Water enters the open pores via diffusion from the ambient medium, the pore pressure drops, and the strength increases. At the third stage, the diffusion rate being higher than the dilatancy rate, water migrating from surrounding rocks is supposed to refill the dilatancy region, restoring various physical parameters to their initial values. The water filling and the pore pressure rise reduce the effective confining pressure, which is equivalent to a decrease in the rock strength. As a result, the growth of tensile cracks accelerates and they join to form a shear rupture resulting in an earthquake. An advantage of the DD model is that it provides a mechanism underlying the dependence of the long-term precursor duration on the strength of an earthquake. A significant drawback of this model is the absence of a mechanism by which the accumulated dilatancy cracks join to form the major rupture.

Dobrovolsky [1980] was first to formulate ideas of a general model containing a heterogeneity (inclusion) that was later called the *consolidation model* [*Dobrovolsky*, 1991]. The geological medium in this model has a block structure, and the earthquake preparation process consists of three stages. A specific feature of the first stage (the regular state) is continuous deformation on a regional or global scale resulting in relative motions of blocks of various scales. Continuous deformation, short-lived bonds between neighboring blocks and their breakage give rise to background seismicity and numerous asynchronous variations in various geophysical fields. A considerable portion of the energy of tectonic processes is spent on gradual deformation of the medium in the regular state phase.

The next phase is consolidation. This phase starts with the formation of a bond between two blocks. If such a bond is sufficiently strong, it will exist for a longer time compared to background bonds. Strain rates will decrease in its vicinity, and conditions for enlargement of the bond area will be created. As a result, the link between the blocks becomes stronger and other blocks can join the first two, forming a consolidated region, or a consolidated heterogeneity.

Relative movements of blocks inside this region are significantly smaller, thereby reducing the seismicity. Note that the existence of the consolidation process is not based on direct experimental data. The formation and development of the consolidated inclusion disturbs the pattern of initial motions, which in turn disturbs the regular pattern of background variations in various geophysical fields. Mechanical stresses and strain rates are redistributed, accommodating movements taking place far from the consolidation region, in which the elastic potential energy is accumulated.

The set of phenomena associated with the consolidation phase is classified as long-term precursors. Their occurrence time is controlled by the intensity of tectonic processes and the growth rate of the consolidated region. The consolidation phase can end with the development of plastic motions on inner interblock boundaries in the consolidated boundary or with a swarm of weak earthquakes. As a result, the accumulated elastic potential energy will decrease. If the decrease in the accumulated energy level is sufficiently large, precursors of a strong earthquake discovered in the consolidation phase will prove false.

The third phase is a fracture phase. At a certain stage, the consolidated region stops growing. This can be related to an unfavorable redistribution of the strain rate field or the influence of adjacent large consolidated blocks. The general geotectonic setting plays a decisive role in this case. The consolidated region is a specific crustal volume whose fracture gives rise to an earthquake. The fracture phase is distinguished by a breakup of the consolidation region into its separate constituent elements (blocks). *Dobrovolsky* notes that the fracture itself occurs mainly on weakened boundaries of blocks, and a rupture can develop in these areas in accordance with various models.

The fracture phase divides into α - and β -stages. At the α -stage, the consolidated region breaks up via plastic movements, foreshocks and major rupture. Short-term precursors are observed at this stage. The time interval from the main shock to the end of the fracture phase is called the β -stage. At this time, the consolidation region is additionally fractured through aftershocks and plastic motions. The total elastic potential energy of the medium drops and the intensity of the aftershock activity slowly decreases. The medium recovers the phase of regular state and the seismic cycle is accomplished.

A significant advantage of the consolidation model is the possibility of obtaining numerical estimates of the strain field and related precursors.

The development of a preparation model including a heterogeneity was dictated by the need for a more general description of the preparation process ignoring various details specific to the AUF, DD, *Stuart*, and *Brady* models. However, this approach has not clarified the formation mechanism of the major rupture and conditions favoring its development. There is every reason to believe that the breakup of the consolidated region complies with the AUF model.

Each of the models presented above has its advantages and drawbacks that apparently set limits to their possible applications to the interpretation of observed experimental data. As noted in [*Zubkov et al.*, 1980], none of the models can be regarded as preferable. Each of these

schemes is acceptable and realizable under appropriate conditions [Meredith *et al.*, 1990]. However, in my opinion, the functional treatment of fracturing in its transition from one structural level to another and the inherent mechanism of a self-developing process, proposed in the AUF model [Myachkin, 1978], makes this model preferable to the other preparation models.

Unique laboratory experiments on the simulation of the preparation process in a macrofracture source have been conducted over the last 10–15 years [Ponomarev, 2003; Sobolev and Ponomarev, 2003]. The AUF model was carefully tested in these experiments, and the majority of prognostic features observed during preparation and development of a macrofracture (earthquake source under conditions of a seismically active region) were experimentally confirmed. These are the division of the loaded medium into zones of different stress levels with essentially different evolutionary patterns of the seismic process, specific behavior of the velocity ratio V_p/V_s and the recurrence plot slope γ , localization of seismicity in a narrow zone defining the position of the future macrofracture, and the clustered fracture of the macrofracture zone. On the other hand, analyzing several precursors of various types, Dobrovolsky arrived at the following conclusion important for the subject of this paper: progress in solving the earthquake prediction problem requires “comprehensive joint analysis of a complex of premonitory phenomena” [Dobrovolsky, 1991].

Choice of precursors. Numerous publications of the 1960s–1990s devoted to diverse premonitory phenomena have provided evidence on about 1000 cases of anomalous behavior of various geophysical fields preceding strong earthquakes. In accordance with physical nature, occurrence time, and variability in space and time, all precursors can be subdivided into several groups of different sizes [Sidorin, 1992; Sobolev, 1993; Zubkov, 2002]. Analysis of the entire set of available data on precursors indicates that the vast majority of precursors are phenomenological, i.e. they simply establish an observation of a geophysical field anomaly before a given strong seismic event. The pertinent publications usually do not present data on statistical characteristics of the precursors because the recorded anomalies are not a result of systematic regular observations; possible, physically substantiated mechanisms responsible for the occurrence of precursors are not considered. As regards the precursor occurrence area, data of observations at individual points (stations) are usually published.

In order to use a concrete precursor for prediction purposes, its significance, i.e. the probability that its occurrence before a strong earthquake was nonrandom, should be estimated using either *a priori* information or retrospective experience. Moreover, it is necessary to develop algorithms for declaring an alarm and estimating the probability of false alarms and missing targets [Keilis-Borok *et al.*, 1980]. In this respect, we should note that only a small number of precursors recorded throughout the world have met the formal criteria developed by the IASPEI subcommission on earthquake prediction [Wyss, 1991].

The above considerations and the experience of the author in developing algorithms for the identification of prog-

nostic anomalies suggest that the choice of precursors usable for the medium-term prediction of earthquakes based on a complex of prognostic criteria should meet the following requirements:

- clear physical meaning of a prognostic criterion;
- physically substantiated relation of each prognostic criterion to the earthquake preparation process;
- availability of observation data for each prognostic criterion; these data should be representative in time (long-term series of prognostic parameter values) and in space (the possibility of their mapping);
- availability of a formal procedure for the identification of anomalies in prognostic parameters based on a model of their behavior during the earthquake preparation;
- possibility of obtaining estimates for retrospective statistical characteristics of each precursor: probability of successful prediction (probability of detection), probability of a false alarm, prognostic efficiency (informativeness), and so on.

Physically substantiated precursors. All prognostic criteria described below belong to the group of seismological parameters estimated from regular seismological observations. Their values are obtained from data processing of catalogs of earthquakes in seismically active regions under study.

Density of seismogenic faults K_{sf} . From the standpoint of the kinetic theory of strength, fracture of solids is a complex thermal fluctuation process extended in time. The endurance of a loaded solid is described by Zhurkov’s equation [Zhurkov *et al.*, 1980]. The stage of final macrofracture of a loaded body is preceded by the stage of formation and accumulation of microcracks with a characteristic lifetime determined by this equation. If l is an average distance between the centers of microcracks that are chaotically distributed in the sample volume, we have $l = c^{-1/3}$, where c is the concentration of cracks. On a given structural scale, this distance can be expressed through the crack size L as follows:

$$l/L = c^{-1/3}/L = K^* . \quad (1)$$

The parameter K^* is known as the concentration criterion of crack enlargement. Its physical meaning consists in the fact that it characterizes the closeness of cracks to each other and thereby their interaction and coalescence ability. The value K^* was found to be virtually constant immediately before macrofracture for wide ranges of materials (rocks included) and crack sizes (from microfractures to earthquake-related ruptures in the crust). According to various estimates of this parameter from data of laboratory and *in situ* experiments, the K^* value range from 2.5 to 6.5.

Based on analogies between fracture processes at various scale levels, the method of concentration criterion of crack enlargement was applied for the first time to large-scale crustal processes in [Sobolev and Zavyalov, 1980] and

later in [Kuksenko *et al.*, 1984]. These studies showed that strong earthquakes mostly occur in areas of lower values of the concentration criterion. For this reason, the concentration parameter of seismogenic faults (fractures) K_{sf} was introduced in seismological practice; this parameter characterizes the state of the seismogenic process in a seismically active crustal volume V_0 at a time moment t .

The fracture concentration parameter K_{sf} has a clear physical meaning and is the ratio of the average distance between seismogenic fractures in the volume V_0 to their average length:

$$K_{sf} = \frac{R_{sf}}{l_{sf}} = \frac{\mu^{-1/3}}{l_{sf}}, \quad (2)$$

where $\mu = N/V_0$ is the volume density (concentration) of fractures determined by earthquakes that occurred in the volume, $l_{sf} = 1/N \sum_j l_j$ is the fault length average over the ensemble of fractures, N is the total number of earthquakes of the energy class range $[K_{min}, K_{max}]$ that occurred in the volume V_0 over the time ΔT , and l_j is the length of an individual seismogenic fault.

Results of analysis of K_{sf} spatiotemporal distributions and the experience of retrospective prediction [Zavyalov, 1986] suggest the following scheme of its application to prediction of strong earthquakes.

(i) The distribution of K_{sf} is calculated with a step of Δt for earthquakes of a given range of energy classes and hypocentral depths in preliminarily chosen elementary seismically active volumes V_i .

(ii) If values $K_{sf} \leq K_{sf}^{al}$, where K_{sf}^{al} is an alarm level determined experimentally from analysis of retrospective data, are found to be present in a certain volume V_i , the strong earthquake expectation period is declared for this volume. The critical level K_{sf}^{al} depends on the seismic activity in the region studied, sizes of the chosen seismically active region, and energy classes of earthquakes involved in the calculations of K_{sf} .

(iii) The expectation state in the volume under consideration V_i is preserved during the time period $\tilde{T}_{exp} + |\sigma_T|$ determined experimentally from retrospective data. This state can end after a strong earthquake whose hypocenter is located in the given seismically active region, whereas in adjacent volumes the expectation state ends with the onset of aftershock activity.

In the first case, the alarm is regarded as real and the time from the beginning of the expectation period (the anomaly onset) to the earthquake moment is defined as the expectation time. In the second case, the alarm declared for adjacent cells that do not contain the hypocenter is considered false (because the place of the event formally was not predicted), and the time from the anomaly onset to the beginning of the aftershock activity is defined as the false alarm period.

(iv) If a strong earthquake does not occur over the time $\tilde{T}_{exp} + |\sigma_T|$ in the volume V_i , this time is classified as the false alarm period.

A typical plot of K_{sf} versus time before a strong earthquake is shown in Figure 1a.

Recurrence plot slope γ . This parameter (b -value) when it is used as a characteristic of earthquake magnitude)

is related to the distribution of earthquakes over their energies (magnitudes) and is one of main characteristics of the seismic regime. The distribution of elastic impulses over their energies in processes of deformation and brittle fracture of various materials and rocks was widely studied under laboratory conditions and in *in situ* experiments (deformations of rock masses in mines) [Mogi, 1962, 1967; Scholz, 1968a, 1968b; Vinogradov, 1957, 1959, 1963, 1964; Zhurkov *et al.*, 1977]. As was shown in a number of studies, this distribution is similar in shape to the distribution of earthquakes over energies. This inference served as an argument for the principle of fracture similarity at various scale levels [Kostrov, 1975; Myachkin, 1978; Myachkin *et al.*, 1975] and provided a basis for the study of the earthquake preparation process under laboratory conditions and in mines [Vinogradov, 1957, 1963]. The validity of this approach was confirmed in [Ponomarev *et al.*, 1997; Smirnov *et al.*, 1995a, 1995b, 1995c; Zavyalov and Sobolev, 1988].

Acknowledging the fundamental nature of the recurrence law, which was initially inferred experimentally, researchers have repeatedly attempted to derive it theoretically [Gurevich *et al.*, 1960; Hausner, 1955; Mogi, 1962; Petrov, 1981]. In all these papers, the authors postulated certain initial presumptions and obtained dependences of the number of earthquakes on their magnitude that were similar to the recurrence law. The theoretical recurrence plot slope in these relations was a constant value.

Pod'yapolsky [1968] and Kuznetsova [1976] arrived at a conclusion, important from the standpoint of prediction studies: the value γ is not constant but depends on material properties and conditions of a loading process. This implies that variations in γ with time observed in laboratory experiments and *in situ* observations yield evidence on the state of material in a fracture source. This idea was further developed and confirmed by Meredith *et al.* [1990].

Laboratory studies of rock fracture and observations in mines have shown that the macrofracture preparation process is characterized by regular temporal variations in the ratio of the numbers of strong shocks to weak ones, which is equivalent to variations in the slope of the plot showing the distribution of acoustic impulses over energies. Studies in mines have shown that, as compared with quiescent periods, the ratio of the numbers of strong pulses to weak ones increases by three to seven times 15–18 hours to 1 month before a rockburst, after which this ratio drops [Cai *et al.*, 1988; Ponomarev *et al.*, 1997; Scholz, 1968a, 1968b; Smirnov *et al.*, 1995a, 1995b, 1995c; Vinogradov, 1957, 1963, 1964]. Similar variations in γ were also observed in detailed studies of seismic regimes in various seismically active regions of the world [Adams, 1976; Katon *et al.*, 1981; Mamadaliev, 1964; Mikhailova, 1980a; Van Wormer *et al.*, 1976; Zavyalov, 2003b; Zavyalov and Sobolev, 1980; Zhang and Fu, 1981].

Thus, there is every reason to suggest that a parameter indicating a forthcoming macrofracture is an anomalous variation in the recurrence plot slope whose value first anomalously increases and then drops until the onset of macrofracture (rockburst or earthquake). As regards the physics of the macrofracture process, the γ increase is associated with the formation of small fractures (weak earthquakes) that are much more numerous compared to the number of large frac-

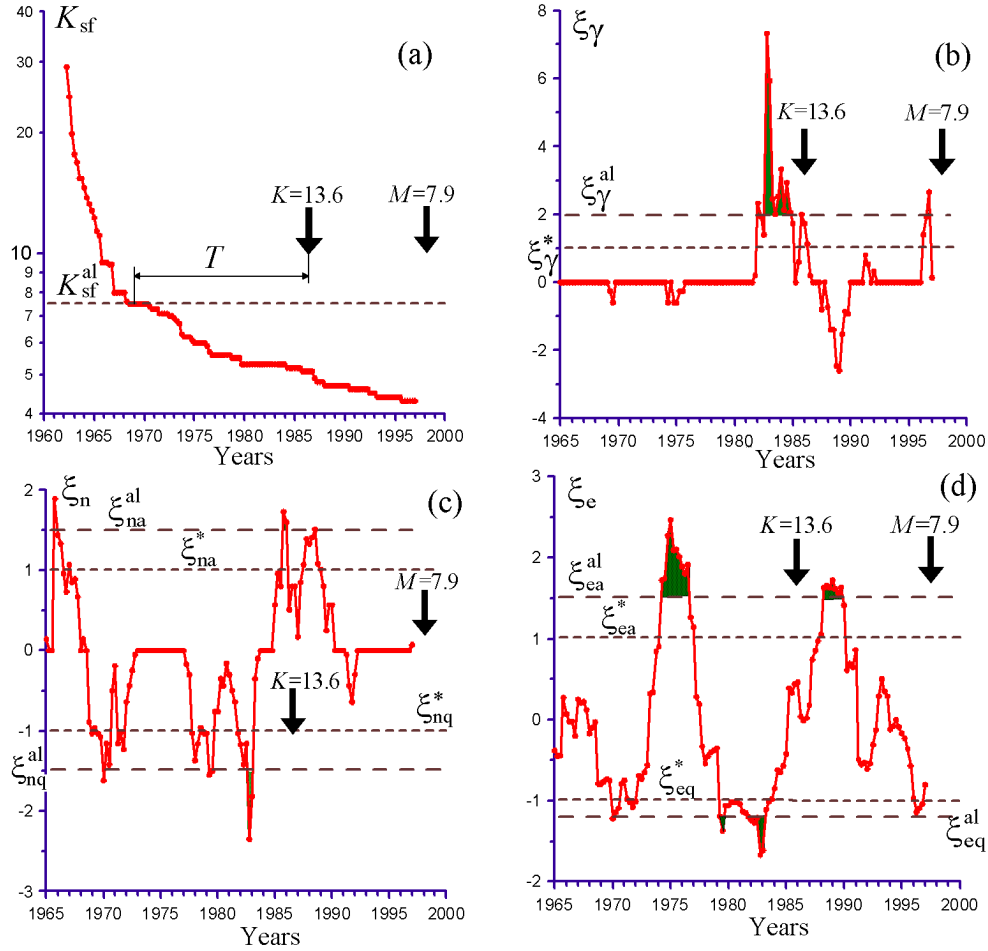


Figure 1. Behavior of precursory parameters in a 100×100 -km cell centered at the December 5, 1997, $M = 7.9$, Kronotski (Kamchatka) earthquake. Each ordinate axis plots a given parameter normalized to the rms error of the determination of its long-term value: (a) density of seismogenic faults K_{sf} ; (b) recurrence plot slope; (c) number of earthquakes per year; (d) released seismic energy $\sum E^{2/3}$. The arrows indicate time moments of strong earthquakes that occurred in the given cell.

tures (moderate earthquakes) in the preparation region of a forthcoming strong earthquake. When the concentration of small fractures reaches a critical value, processes of interaction between small cracks and their coalescence become active; as a result, the number of large fractures increases relative to small ones and the value of γ decreases until the onset moment of the major rupture (a strong earthquake).

Using the maximum likelihood method, the value γ is calculated by the formula [Gusev, 1974]

$$\gamma = \lg \left[1 + \frac{N_\Sigma}{\sum_{n=0}^{\infty} n \cdot N(K_{\min} + n \cdot \Delta K)} \right] / \Delta K, \quad (3)$$

$$\sigma_\gamma \approx \gamma / \sqrt{N_\Sigma}. \quad (4)$$

Here N_Σ is the total number of earthquakes in the range of energy classes from K_{\min} to K_{\max} (the sample size); $N(K_{\min} + n \cdot \Delta K)$ is the number of earthquakes of the energy class $K_{\min} + n \cdot \Delta K$, $n = 0, 1, 2, \dots$, and ΔK is the unity of the energy class range.

Using the parameter γ for prediction analysis of strong earthquakes, we represented its temporal variations in terms of a dimensionless quantity ξ_γ characterizing the statistical significance of deviations of current values of the parameter γ from its long-term values normalized to the rms error of their determination (7).

The following formal rule was used for the identification of precursory anomalies in the parameter ξ_γ [Zavyalov, 1984b]. The criterion of the anomaly onset is the increment in ξ_γ above a certain alarm level ξ_γ^{al} chosen empirically from a retrospective examination of available data. Anomalous values of ξ_γ reach a maximum followed by their decrease, sometimes to negative values. The termination moment of the ξ_γ

anomaly is defined by its transition through a level ξ_γ^* determined also empirically from a new increase in the ξ_γ value. The time from the onset of the anomaly to its termination is regarded as its period T_{an} .

The anomaly is considered as realized if a single earthquake of the predicted energy class K_{pr} or a group of such earthquakes occurs within the anomaly period T_{an} in the seismically active subregion (cell) under study ΔS_i . In this case, the time from the anomaly onset to the moment of the seismic event is the expectation time of this event T_{exp} (evidently, $T_{\text{exp}} \leq T_{\text{an}}$). If no strong earthquake occurs during the time T_{an} , the anomaly is considered as false and its duration is the false alarm time $T_{\text{f.a.}}$.

The above model describing the behavior of the precursor ξ_γ is consistent with results of laboratory experiments on fracture of rocks and model materials, as well as with the AUF model concepts.

The following formal scheme implementing the use of the parameter ξ_γ for prediction of strong earthquakes was proposed in [Zavyalov, 1984b].

(i) Spatial distributions of the parameter ξ_γ are calculated, at step Δt , for preliminarily chosen cell sizes ΔS_i and widths of the current (ΔT_c) and long-term (ΔT_l) moving time windows.

(ii) As soon as values $\xi_\gamma \geq \xi_\gamma^{\text{al}}$ appear in a subregion ΔS_i , the expectation period of a strong earthquake is declared in this subregion.

(iii) The expectation state is preserved for the time period $\tilde{T}_{\text{exp}} + |\sigma_T|$ determined empirically from retrospective analysis of available data and can be terminated before the expiration of the expectation period in the following two cases:

(a) after the occurrence time of a strong earthquake in the subregion considered;

(b) with a new increase in the value ξ_γ above the level $\xi_\gamma^* = +1$, following the period of its decrease.

In the first case, the time from the anomaly onset to the strong earthquake occurrence is the expectation time T_{exp} . The time interval in the second case is the false alarm period $T_{\text{f.a.}}$. A typical dependence of ξ_γ versus time before a strong earthquake is shown in Figure 1b.

Number of events per unit time N. A very important characteristic of rock fracture under laboratory conditions and of the seismic regime in a seismically active region is the number of acoustic pulses (earthquakes) recorded per unit time. *Vinogradov* [1980] noted that the time distribution of elastic impulses characterizes the evolution of fracture processes in the study volume depending on the strain rate, material properties and, primarily, heterogeneity of the medium. *Sobolev* [1993] pointed out that the acoustic emission induced by deformation of rock samples reflects the fracturing process and can be considered, with some reservations, as an analogue of the seismic process in the Earth.

A review of published data on the generation of acoustic (seismic) impulses indicates that the study of the temporal distribution of elastic impulses (background earthquakes in seismically active regions) is beneficial to prediction of a macrofracture (an earthquake) [Atkinson, 1981; Borovik, 1977; Evisson, 1977; Fedotov et al., 1969; Inouye,

1965; Kerr, 1979; Mikhailova, 1980a; Mogi, 1962, 1968a, 1968b; Nersesov et al., 1976, 1979; Rummel and Sobolev, 1983; Shamina, 1956; Teitelbaum and Ponomarev, 1980; Tomashevskaya and Khamidullin, 1972; Vinogradov, 1957, 1963; Vinogradov et al., 1975; Zavyalov and Sobolev, 1980, 1988]. Strong earthquakes can be preceded by periods of both attenuation of seismic activity (a decrease in the number of background earthquakes) and its enhancement (an increase in the number of background earthquakes). In some cases, both types of the premonitory variations separated in time can be observed in the same region. These schemes are qualitatively accounted for in terms of the AUF model of earthquake preparation.

The average number of earthquakes per unit time in the range of energy classes from K_{min} to K_{max} and its rms error are determined by the formulas

$$N = \frac{1}{\Delta T} \sum_{K_{\text{min}}}^{K_{\text{max}}} N(K), \quad \sigma_n = \sqrt{N}, \quad (5)$$

where $N(K)$ is the number of earthquakes of the energy class K , and ΔT is the length of the observation period (time window width). A dimensionless parameter ξ_n defined in the same way as the parameter ξ_γ is taken as the measure of statistical significance.

If *seismic quiescence* is sought, the anomaly onset time is determined from the appearance of values $\xi_n \leq \xi_{nq}^{\text{al}}$, where ξ_{nq}^{al} is an empirically chosen alarm level. If *seismicity activity* is sought, the anomaly onset time is determined from the appearance of values $\xi_n \geq \xi_{na}^{\text{al}}$, where ξ_{na}^{al} is an empirically chosen alarm level. The rule for the anomaly identification from the parameter ξ_n is similar to the rule described above for the parameter ξ_γ and complies with the generally acknowledged idea that an earthquake is preceded by either seismic quiescence or a foreshock activation period.

An anomaly is considered false if no earthquakes of predicted energy class K_{pr} occurs over the time $\tilde{T}_{\text{exp}} + |\sigma_T|$ in the seismically active volume considered. In this case, the anomaly duration is the false alarm period. On the other hand, if an earthquake or a group of earthquakes of the predicted energy class occurs in the time interval $\tilde{T}_{\text{exp}} + |\sigma_T|$, the time from the anomaly onset to the earthquake occurrence is regarded as the expectation period for the given seismic event and, with the end of this period the anomaly is regarded as realized.

The scheme of using the parameter ξ_n for prediction of strong earthquakes is similar to that proposed for the parameter ξ_γ . A typical example of the dependence of ξ_n versus time before a strong earthquake is given in Figure 1c.

Released seismic energy $\sum E^{2/3}$. The parameter $\sum E^{2/3}$ was proposed and tested by *Keilis-Borok* [1959] for strong ($M > 7$) earthquakes. The physical meaning of this parameter consists in rapid accumulation of dynamic rupture areas producing earthquakes smaller than a given energy class K_0 in the preparation region of a strong earthquake [Sadovsky and Pisarenko, 1983]. *Keilis-Borok and Malinovskaya* [1966] noted the following important relation: a strong earthquake occurs when the sum of the weak earth-

quake rupture areas becomes equal to the area of the main rupture. According to the AUF model of earthquake preparation, the dependence $E(t) = \sum E^{2/3}$ plotted for weak earthquakes over time intervals ΔT much shorter than the preparation time of the main rupture should have a baylike shape in the fracturing process [Myachkin, 1978; Myachkin et al., 1975]. Myachkin et al. [1975] emphasized that such behavior of a precursor is only possible immediately in the zone of the avalanche-like growth of fractures and the formation of the future main rupture, which is important for predicting the place of the forthcoming shock.

The parameter $\sum E^{2/3}$ is determined by the formula

$$\sum E^{2/3} = \frac{1}{\Delta T} \sum_{i=1}^{N_{\Sigma}(K < K_0)} E_i^{2/3}, \quad (6)$$

where $N_{\Sigma}(K < K_0)$ is the number of earthquakes of energy classes smaller than the given level K_0 .

Similar to the cases of the recurrence plot slope and the number of earthquakes considered above, the dimensionless parameter ξ_e (7) is accepted as the measure of statistical significance. The released seismic energy, similar to the number of earthquakes, is used in two modifications: as *energy quiescence* (the parameter ξ_{eq}) and as *activation of seismic energy release* (ξ_{ea}). The scheme of the identification of precursory anomalies from the parameter ξ_e and its application to prediction of strong earthquakes is similar to those described above for ξ_{γ} and ξ_n . A typical dependence of ξ_e versus time before a strong earthquake is shown in Figure 1d.

The above can be summarized as follows. A short review of earthquake preparation models has revealed their advantages and drawbacks setting limits to their possible applications to the interpretation of observed experimental data and the development of earthquake prediction methods. A general characteristic of these models consists in the fact that they are qualitative and useless for a quantitative description of the fracture preparation process in heterogeneous media including deep rocks of the Earth. An exception is the consolidation model allowing one to gain numerical constraints on the strain field and related precursors.

The preparation models, describing the behavior of prognostic parameters at various stages of the earthquake preparation process, provided a physical basis for the medium-term earthquake prediction from a complex of prognostic parameters and pointed out requirements to the relevant choice of precursors. Prognostic parameters meeting these requirements and included for their use in the algorithm of mapping expected earthquakes (MEE) were described.

Based on physical principles and the experience of retrospective analysis using the criteria described above, we developed models describing the behavior of prognostic parameters during preparation of earthquakes and proposed algorithms for their use in prediction practice.

4. Prognostic Algorithm of Mapping Expected Earthquakes (MEE)

The MEE algorithm is a complex of procedures designed for estimating the conditional probability of strong earthquake occurrence $P(D_1|K)$ in a time period $[t_0, t_0 + \Delta T_{MEE}]$ in rectangular cells of given sizes that form a grid covering the seismically active region studied. A map of expected earthquakes is meant as the set of conditional probability values $P(D_1|K)$ calculated for all spatial cells in the given region and represented in the form of contours. The value ΔT_{MEE} denotes the serviceable time of the map of expected earthquakes (EE) during which a strong earthquake can occur.

We constructed the MEE algorithm using the Bayes approach as the most widespread and simple technique. It is based on the use of probabilities that certain indicators (prognostic parameters K_i) assume values complying with a diagnosis (hypothesis) D_j .

Main principles of the MEE algorithm construction. The MEE algorithm of earthquake prediction is based on the principle of spatial-temporal scanning of a catalog of earthquakes in the seismically active region to be analyzed. The region is covered by a rectangular grid of $\Delta X \times \Delta Y$ cells, and values of various prognostic parameters are calculated for each cell in a moving time window of a width ΔT at a step Δt . Neighboring cells are either independent or can partly overlap each other [Sobolev et al., 1990, 1991; Zavyalov et al., 1995; Zavyalov, 2002]. Overlapping of neighboring cells can be used to partly estimate uncertainties of epicenter determinations.

The main database used in the MEE algorithm is a catalog of earthquakes in the seismically active region under study; this catalog should be representative and homogeneous for earthquakes of energy classes $K \geq K_0 (M \geq M_0)$.

Two types of diagnoses are possible in the case of earthquake prediction: occurrence (diagnosis D_1) or nonoccurrence (D_2) of a strong earthquake in the spatial cell under consideration. The probabilities of these diagnoses $P(D_1)$ and $P(D_2)$ are calculated using the earthquake catalog, without regard for any prognostic parameters; the calculated values (Table 1) are unconditional probabilities (UPs).

The MEE algorithm uses two groups of prognostic parameters, quasi-stationary and dynamic. The first group includes geological and geophysical parameters that do not vary or vary weakly during an observation period and the strong earthquake preparation time, are physically related to the occurrence of earthquakes, and can be mapped throughout the observation area (see Chapter 3). The quasi-stationary parameters are used for differentiating the observation area according to the level of unconditional probability of strong earthquake occurrence $P(D_1)$. Experience of MEE applications in various seismically active regions of the world showed that the incorporation of quasi-stationary indicators the unconditional probability $P(D_1)$ by a factor of 1.5–2 in cells where they are observed and decreases it by the same factor in cells where they are not observed. The following quasi-stationary parameters can be utilized:

Table 1. Retrospective statistical characteristics of prognostic criteria

Unconditional probability of earthquake occurrence	$P(D_1) = \lambda \exp(-\lambda), \lambda = \frac{\tilde{S}_{\text{exp}}^c}{S_{\text{obs}}} \cdot \frac{\tilde{T}_{\text{exp}}^c}{T_{\text{obs}}} \cdot N_{\text{total}},$ S_{obs} and T_{obs} are area and time of observations
Detection probability	$P(K_i D_1) = N_{\text{pr}}/N_{\text{total}}$
False alarm probability	$P(K_i D_2) = S_{\text{f.a.}}/S_{\text{obs}}^*$
Average expectation (false alarm) time in relation to the parameter K_i	$\tilde{T}_{\text{exp(f.a.)}}^i = 1/N_S \sum_{j=1}^{N_S} T_{\text{exp(f.a.)}}^j,$ N_S is the number of cells in the expectation state
Average expectation (false alarm) time in relation to a complex of parameters	$\tilde{T}_{\text{exp(f.a.)}}^c = 1/n \sum_{i=1}^n \tilde{T}_{\text{exp(f.a.)}}^i,$ n is the number of parameters
Average expectation (false alarm) area in relation to the parameter K_i	$\tilde{S}_{\text{exp(f.a.)}}^i = 1/N_S \sum_{j=1}^{N_S} S_{\text{exp(f.a.)}}^j$
Average expectation (false alarm) time in relation to a complex of parameters	$\tilde{S}_{\text{exp(f.a.)}}^c = 1/n \sum_{i=1}^n \tilde{S}_{\text{exp(f.a.)}}^i$
MEE serviceability time	$\Delta T_{\text{MEE}} = \tilde{T}_{\text{MEE}}^c = 1/n \sum_{i=1}^n \tilde{T}_{\text{MEE}}^i$
Average area of alarm zones with a given level of conditional probability	$\tilde{S}_{\text{al}} = 1/N_{\text{MEE}} \sum_{j=1}^{N_{\text{MEE}}} S_{\text{al}}^j,$ N_{MEE} is the number of EE maps
Prognostic effectiveness of a precursor	$J_t = \frac{N_{\text{pr}}/T_{\text{exp.total}}}{N_{\text{total}}/T_{\text{obs}}}, J_S = \frac{N_{\text{pr}}/S_{\text{exp.total}}}{N_{\text{total}}/S_{\text{obs}}}$
Prognostic effectiveness of the MEE algorithm	$J_{\text{MEE}} = \frac{N_{\text{pr}}/\tilde{S}_{\text{al}}}{N_{\text{total}}/S_{\text{obs}}}$

the number of tectonic faults of a given rank in a cell, the number of fault intersections in a cell, the velocity gradient of recent movements, gravity anomalies, etc.

Dynamic criteria include geophysical parameters estimated from regular seismological observations, as well as observations of other geophysical fields that have clear physical meaning and are related to the preparation of strong earthquakes. The variation period of the dynamic parameters is much shorter than the observation period and the preparation time of a strong earthquake. A necessary condition for using a given dynamic criterion in the MEE algorithm is the possibility of its mapping (see the paragraph “Choice of Precursors”).

No limitations are imposed on the number of quasi-stationary and dynamic prognostic parameters used in the MEE algorithm and, in this sense, the algorithm is open.

In regular observations of geophysical parameters, a researcher is often interested not in the values of these parameters but in their statistically significant deviations from their long-term values. The statistical significance of the observed deviations of prognostic parameters is particu-

larly important in establishing their relation to subsequent catastrophic natural phenomena. This approach is used in the MEE algorithm.

All dynamic prognostic parameters K_i , with the exception of the seismogenic fault concentration K_{sf} having a cumulative nature and being a threshold quantity, are represented as spatial-temporal distributions of anomalous deviations from a corresponding long-term (background) level that are normalized to the rms error of its determination. The parameter ξ_{K_i} taken as a measure of statistical significance is defined as follows (Figure 2) [Zavyalov, 1984a, 1984b]:

$$\xi_{K_i} = \begin{cases} \frac{|\Delta K_i - \text{Sign}(\Delta K_i) \cdot |\sigma K_{ic}|}{|\sigma K_{ic}|}, & \text{if } |\Delta K_i| > |\sigma K_{ic}| \\ 0, & \text{if } 0 \leq |\Delta K_i| \leq |\sigma K_{ic}|, \end{cases} \quad (7)$$

where $\Delta K_i = K_{ic} - K_i$; K_{ic} and σK_{ic} are the current value of the prognostic parameter and its standard deviation in the moving time window ΔT_c calculated from pertinent

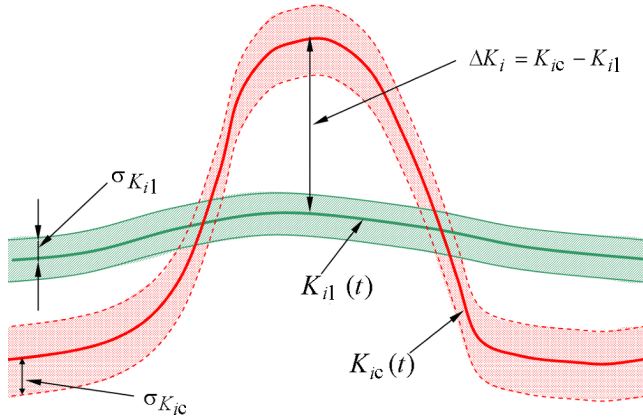


Figure 2. Schematic illustration of the meaning of the parameter ξ_{K_i} .

formulas (see Chapter 3); and K_{i1} and $\sigma_{K_{i1}}$ are long-term (background) values calculated in the moving time window $\Delta T_1 > \Delta T_c$.

Each dynamic parameter has retrospective statistical estimates obtained for chosen alarm levels; the most important of these estimates are the detection probability $P(K_i|D_1)$, i.e. the probability of occurrence of a strong earthquake after anomalous values of the parameter K_i were recorded; the false alarm probability $P(K_i|D_2)$, i.e. the potential of nonoccurrence of a strong earthquake after anomalous values of K_i were recorded; the average time of expectation of a strong earthquake \tilde{T}_{exp}^i ; and the average expectation area \tilde{S}_{exp}^i .

The expectation time T_{exp}^i is defined as the time from the initial moment of recording K_i values that exceed a given alarm level (anomalous values) to the occurrence moment of a strong earthquake. If no earthquake occurs, this time is considered as the false alarm time $T_{f.a.}^i$.

The expectation area S_{exp}^i of a strong earthquake in relation to the prognostic parameter K_i is the sum of areas of $\Delta X \times \Delta Y$ cells in which the expectation state is declared, i.e. K_i values in these cells are anomalous, exceeding the given alarm level. If no strong earthquake occurs in these cells, their total area is considered as the false alarm area $S_{f.a.}^i$.

Expressions used for the determination of retrospective statistical characteristics are presented in Table 1.

The alarm level for each prognostic parameter is chosen through expert analysis based on the set of retrospective statistical estimates and other *a priori* information.

The detection probability $P(K_i|D_1)$ is defined as the ratio of the number of strong earthquakes N_{pr} preceded by an anomaly of the parameter K_i (predicted strong earthquakes) to the total number of strong earthquakes N_{total} that occurred during the observation period in the observation area [Zavyalov *et al.*, 1995]. In turn, the false alarm probability $P(K_i|D_2)$ is the ratio of the false alarm area to the observation area (Table 1).

The conditional probability of strong earthquake occurrence $P(D_1|K)$ determined from a complex of prognostic parameters in each spatial cell ΔS_{MEE} is calculated by the Bayesian formula; given two diagnoses, this formula has the form

$$P(D_1|K) = \frac{P(D_1) \prod_{i=1}^n P(K_i|D_1)}{P(D_1) \prod_{i=1}^n P(K_i|D_1) + P(D_2) \prod_{i=1}^n P(K_i|D_2)}, \quad (8)$$

In the absence of anomalous values of the prognostic parameter K_i , the probabilities $P(K_i|D_1)$ and $P(K_i|D_2)$ in (8) are replaced by the probabilities $1 - P(K_i|D_1)$ and $1 - P(K_i|D_2)$. We assume that all parameters used here are simple, i.e. the presence of anomalous values of K_i in the cell under consideration is coded as “1” and their absence, as “0”. If quasi-stationary prognostic parameters K'_i are used, the unconditional probabilities $P(D_1)$ and $P(D_2)$ in grid cells are replaced by the stationary conditional probabilities $P(K'_i|D_1)$ and $P(K'_i|D_2)$. Values of the latter are calculated by formula (8).

The set of conditional probabilities $P(D_1|K)$ for all spatial cells represented in the form of contours was called the map of expected earthquakes for the time period $[t_0, t_0 + \Delta T_{MEE}]$, where ΔT_{MEE} is the MEE serviceability time (Table 1). Occurrence of a strong earthquake in this time interval is assumed to be equiprobable.

The area of MEE alarm zones S_{exp}^i is the total area of cells in which the conditional probability of strong earthquake occurrence $P(D_1|K)$ exceeds a given level. The average area of alarm zones \tilde{S}_{al} with a given level of conditional probability of strong earthquake occurrence is determined by the expression given in Table 1.

In order to select prognostic criteria for their use in the calculation of the conditional probability $P(D_1|K)$, we estimated their prognostic effectiveness in time (J_t) and in area (J_s). The prognostic effectiveness of each criterion was estimated retrospectively, as the ratio of the average flow density of strong earthquakes over the alarm time (or over the alarm area) to their average density over the observation time (or over the observation area) (Table 1) [Fedotov *et al.*, 1976]. Here, N_{pr} and N_{total} are, respectively, the number of predicted strong earthquakes and the total number of earthquakes that occurred over the observation time T_{obs} in the observation area S_{obs} ; $T_{exp, total}^j$ is the total time of expectation (alarm); and $S_{exp, total}^j$ is the total area of expectation (alarm). This definition implies that $J = 1$ if a successful prediction was random (the prognostic criterion is unrelated to the preparation of a strong earthquake), when the densities of earthquakes to be predicted in the alarm and observation periods (areas) coincide. The larger the deviation of the effectiveness parameter from 1, the more effective the prognostic criterion in use. Prognostic criteria with an effectiveness of less than 1.2 were rejected from MEE calculations.

The prognostic effectiveness of the MEE algorithm as a whole, J_{MEE} (Table 1), is defined as the ratio of the flow density of strong earthquakes in the average alarm area to

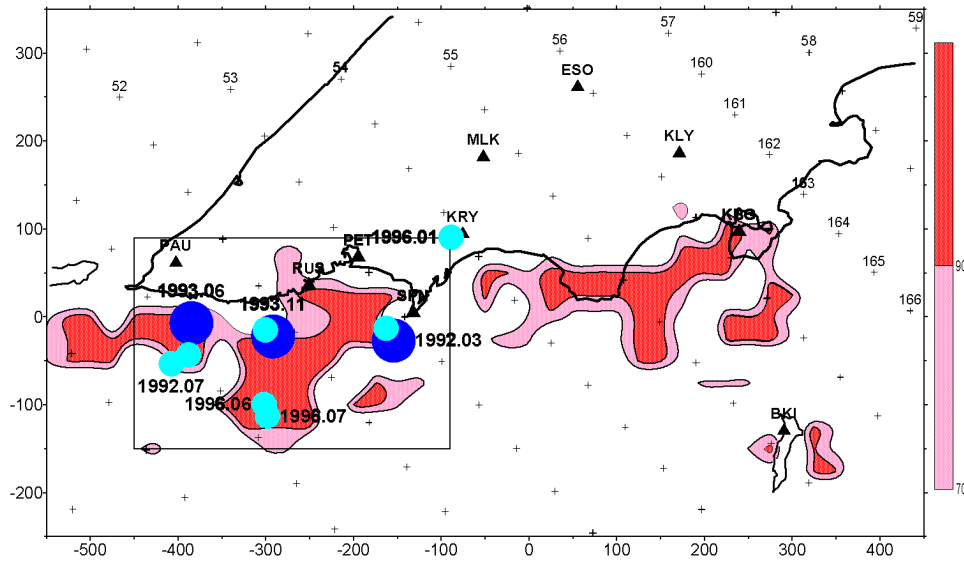


Figure 3. Example of a Kamchatka EE map for the period 1992–1996 calculated from earthquake catalog data over 1962–1991. Epicenters of the $K \geq 13.5$ earthquakes of 1992–1996 are shown in the map as circles proportional in size to the source rupture length on the map scale. Distances along the X and Y axes are measured in kilometers. The geographic coordinate grid is shown by symbols “+”. The legend shows shadings corresponding to various levels of $P(D_1|K)$.

its value in the observation area at a given level of seismic activity (commonly no less than one earthquake per year).

5. Testing of the Prognostic MEE Algorithm in Various Seismically Active Regions from 1985 through 2002

Initially, the MEE algorithm was developed and tested for the Caucasus [Sobolev *et al.*, 1990, 1991; Zavyalov, 1993]. Afterward it was tested in seismically active regions of Kamchatka [Zavyalov and Levina, 1993]; Turkmenistan [Zavyalov and Orlov, 1993]; Kyrgyzstan, southern California, northeastern and southwestern China [Zavyalov and Zhang, 1993]; Greece [Zavyalov, 1994, 2003a; Sobolev *et al.*, 1999]; and western Turkey [Zavyalov, 2003b]. In all of these regions differing in their seismotectonic conditions, a unified approach was applied to the distribution of prognostic parameters, estimation of their retrospective statistical characteristics, choice of optimal alarm levels and the strength of predicted earthquakes, and so on. In doing so, we proceeded from the assumption on the similarity between earthquake preparation processes in seismotectonically different regions based on the concept of self-similarity of geodynamic processes [Sadovskiy, 1986]. This approach proved effective in accumulating, over a relatively short time interval, statistically rather representative data sufficient in order to draw reliable conclusions concerning prognostic capabilities of the MEE algorithm.

The main database of the MEE algorithm was compiled from regional earthquake catalogs, each including a few thousand events and covered a period of a few tens of years. Depending on a region, earthquakes with $K_{pr} \geq 12.5$ ($M_{pr} \geq 5.5$) were objects of prediction analysis. Stationary

prognostic parameters were used in some seismically active regions.

Comparison of optimal alarm levels chosen in various seismically active regions showed that, for each precursory parameter, these levels vary weakly from region to region. This indicates the stability of the algorithm parameters in relation to various seismically active regions differing in geological and tectonic conditions. For the majority of precursors in all regions studied, the detection probability estimated from each precursor is a few times lower than the conditional probability $P(D_1|K)$ amounting to 70% (90% if a complex of the precursors is used). Usually, the detection probability obtained from an individual precursor is 1.5–2 times higher than the unconditional probability $P(D_1)$ but can occasionally be even lower than the latter. An exception is the parameter K_{sf} yielding a detection probability higher than the unconditional probability by a factor of 2.5 and more. The conditional probabilities of strong earthquake occurrence $P(D_1|K) = 70\%$ and 90% were chosen as alarm levels for the analysis of a complex of prognostic parameters.

The prognostic effectiveness of all parameters varies from region to region. The most effective precursor in all regions is the parameter K_{sf} . Its prognostic effectiveness averages $J_t = 4.96 \pm 2.39$. A parameter next in effectiveness is $N_{act} - J_t = 3.96 \pm 1.66$. Less effective are the quiescence in relation to the number of events ($J_t = 3.19 \pm 0.94$) and the recurrence plot slope ($J_t = 2.24 \pm 0.73$). The two parameters related to the seismic energy release proved least effective. The parameter E_q (quiescence in relation to the energy release) had an overly low effectiveness ($J \leq 1$) in several regions and was not used there.

As an example, the Kamchatka EE map is shown in Figure 3.

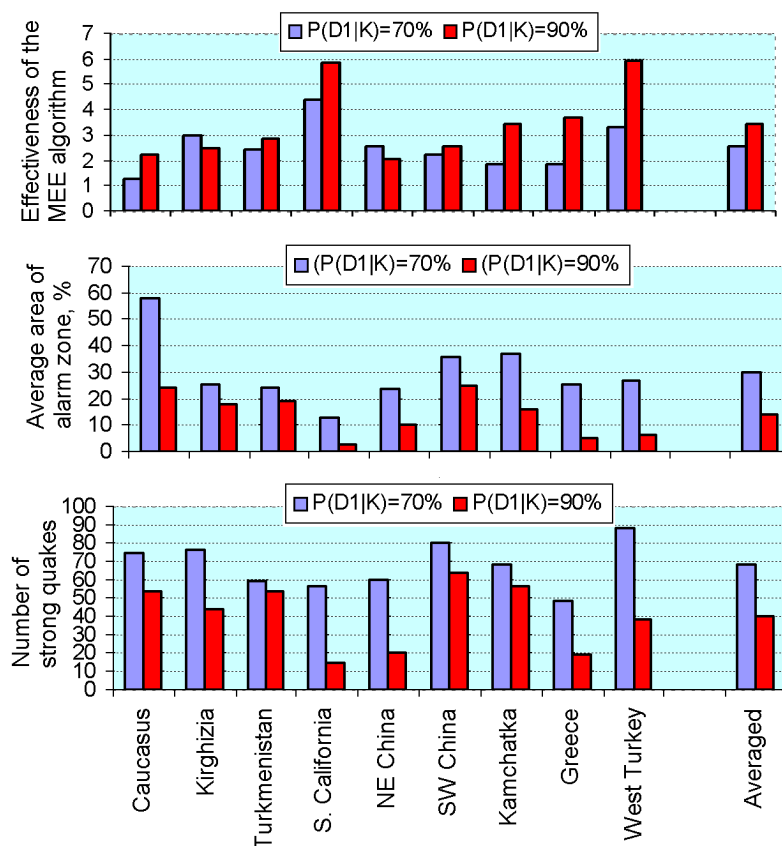


Figure 4. The MEE algorithm effectiveness parameters in various regions: (a) effectiveness of the MEE algorithm J_{MEE} ; (b) average area of alarm zones \tilde{S}_{a1} ; (c) number of earthquakes that occurred in a given alarm zone.

Analysis of a set of EE maps in the study regions have shown that, using the value $P(D_1|K) = 70\%$ as an alarm level (exceeding the unconditional probability in these regions by about three times), the MEE algorithm effectiveness J_{MEE} ranges from 1.29 (the Caucasus) to 4.40 (S. California), whereas J_{MEE} varies within the range 2.02–5.84 with the level $P(D_1|K) = 90\%$. The average value of the MEE algorithm effectiveness J_{MEE} amounts to 2.45 ± 0.94 for the level $P(D_1|K) = 70\%$ and 3.14 ± 1.23 for $P(D_1|K) = 90\%$. In zones with $P(D_1|K) \geq 70\%$, 48% (Greece) to 80% (SW China) of the predicted earthquakes occurred over the serviceability time of the corresponding EE maps (a few years). The average area of the corresponding alarm zones was 12.7% to 58.1% of the observation area with a seismicity level of at least one earthquake per year (commonly, 1.5–2 times smaller than the total area of observations). These data are plotted in Figure 4.

Results of the long-term testing of the MEE algorithm in various seismically active regions showed that its average prognostic effectiveness is 2.5 times higher than the random guessing value. On average, 68% of the predicted earthquakes occurred in zones with $P(D_1|K) \geq 70\%$, and the average area of alarm zones amounted there to 30% of the area with a seismicity level of at least one earthquake per

year. These data yield an idea of the real potential prognostic capabilities of the MEE algorithm.

Figure 3, illustrating a rather typical example, shows that several large zones of high values of $P(D_1|K)$ exist in seismically active regions. Detailed EE maps, with smaller sizes of spatial cells, were constructed for a southern seismically active 360×240 -km region of Kamchatka from Cape Lopatka to Cape Shipunski (the rectangular area in Figure 3). The cell areas amounted to 30×30 km for calculating the parameter K_{sf} and 60×60 km for the remaining precursory parameters. One of the EE maps of this region is presented in Figure 5.

Comparison of retrospective statistical characteristics for the whole Kamchatka Peninsula and its southern part reveals a decrease in the prognostic effectiveness of all precursory parameters in the whole Kamchatka region, which is related to an increase in the false alarm probability. The K_{sf} effectiveness decrease is particularly pronounced (by more than two times). However, the alarm level and the average expectation time changed insignificantly, indicating that the algorithm parameters are stable with respect to variations in the spatial cell sizes. Analysis of the whole series of maps for southern Kamchatka leads to the conclusion that the prognostic effectiveness of the EE map series for southern Kamchatka is more than two times higher than its value for

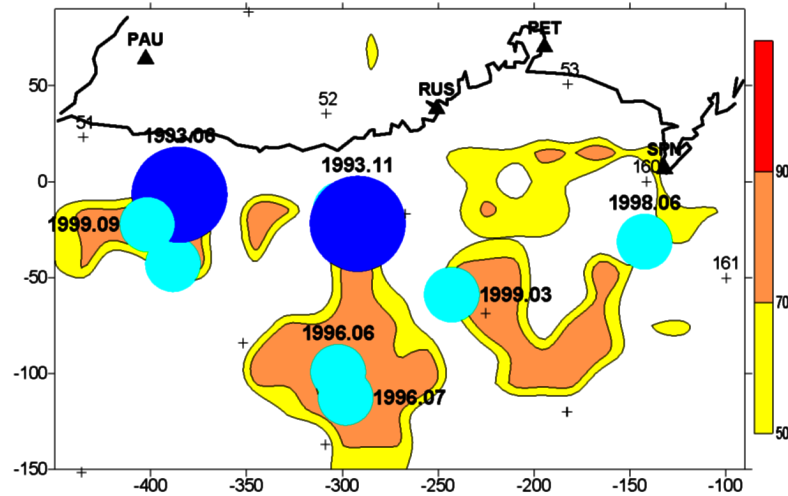


Figure 5. EE map of the Cape Shipunski-Cape Lopatka area (Kamchatka) for the period 1993–1999. The map shows the epicenters of $K \geq 13.5$ earthquakes that occurred during the expectation period. The rest of notation is the same as in Figure 3.

the whole Kamchatka region. This appears to be primarily related to a decrease in the alarm area in this part of the region.

In order to examine the effectiveness of the MEE algorithm as a function of the energy class of predicted earthquakes and the depth range of the seismically active layer, we calculated EE map series for Kamchatka with the following parameters: $K_{pr} \geq 14.5$, $H = 0-100$ km; $K_{pr} \geq 13.5$, $H = 0-50$ km (Figure 6a); and $K_{pr} \geq 13.5$, $H = 25-75$ km (Figure 6b). Retrospective statistical characteristics in these cases also display a low sensitivity of the algorithm parameters to variations in earthquake parameters involved in the calculations. Analysis of the EE maps constructed in these three variants shows that their prognostic effectiveness is higher compared to the basic variant ($K_{pr} \geq 13.5$, $H = 0-100$ km). A high efficiency of the $K_{pr} \geq 14.5$ earthquake prediction ($J_{MEE} = 6.5$ with $P(D_1|K) = 70\%$) is due to a significant decrease in the alarm area. Because of small statistics of $K_{pr} \geq 15.5$ earthquakes (only three such events occurred over the period from 1962 through 2000), this energy range of predicted earthquakes was not analyzed.

The maps presented in Figures 6a and 6b show that the epicentral positions of the main shock of the Kronotski, December 5, 1997 earthquake ($K = 15.5$, $H = 10$ km) and several of its aftershocks coincide with the zone of $P(D_1|K) \geq 70\%$, whereas the larger part of the aftershock area is characterized by values $P(D_1|K) \geq 50\%$. On the other hand, as seen from Figure 6b showing the distribution of the conditional probability $P(D_1|K)$ for the depths $H = 25-75$ km, only an insignificant part of the aftershock zone has its values in the range $50\% \leq P(D_1|K) \leq 70\%$. Such a distribution pattern implies that the preparation process of the Kronotski earthquake developed mostly in the depth range $H = 0-25$ km, which agrees with data of instrumental determinations of its hypocentral depth. Apparently,

the MEE algorithm is capable of predicting the source depth of a future earthquake.

Calculations of EE maps for western Turkey with the cell sizes $\Delta S_{MEE} = 50 \times 50$ km, 75×75 km and 100×100 km provided constraints on the effect of the averaging cell size on the prognostic properties of the MEE algorithm. The MEE algorithm proved most effective with the cell size $\Delta S_{MEE} = 75 \times 75$ km ($J_{MEE} = 3.29$), whereas its effectiveness was lowest in the case of $\Delta S_{MEE} = 50 \times 50$ km ($J_{MEE} = 1.19$). We may suggest that calculations with various sizes of averaging cells and comparison of their results with estimated sizes of the earthquake source and preparation areas as a function of the earthquake magnitude will be helpful for more accurate adjustment of the MEE algorithm to earthquake prediction in a given range of magnitudes.

The long-term testing of the MEE algorithm has revealed some of its drawbacks. The most significant of the latter is as follows. As mentioned above, EE maps usually exhibit zones of high conditional probabilities of the occurrence of a strong earthquake. Several alarm zones are also observable even in the EE maps constructed for the Cape Lopatka-Cape Shipunski in Kamchatka (Figure 5). This implies that, using solely the prognostic capabilities of the MEE algorithm and related individual precursors, one cannot localize the zone of a future strong earthquake. This can be done by invoking, first, shorter-term precursors (compared to precursors involved in the MEE algorithm procedure) and, second, precursors of the nonseismological type (e.g. the water level in wells, radon concentration, electromagnetic emission, and so on). Therefore, the MEE algorithm can be recommended, on the one hand, as a means complementing observations of other precursors in inferred zones of high conditional probabilities (more than 70%) and, on the other hand, for taking preventive measures to mitigate the possible economic and social damage of a future strong earthquake.

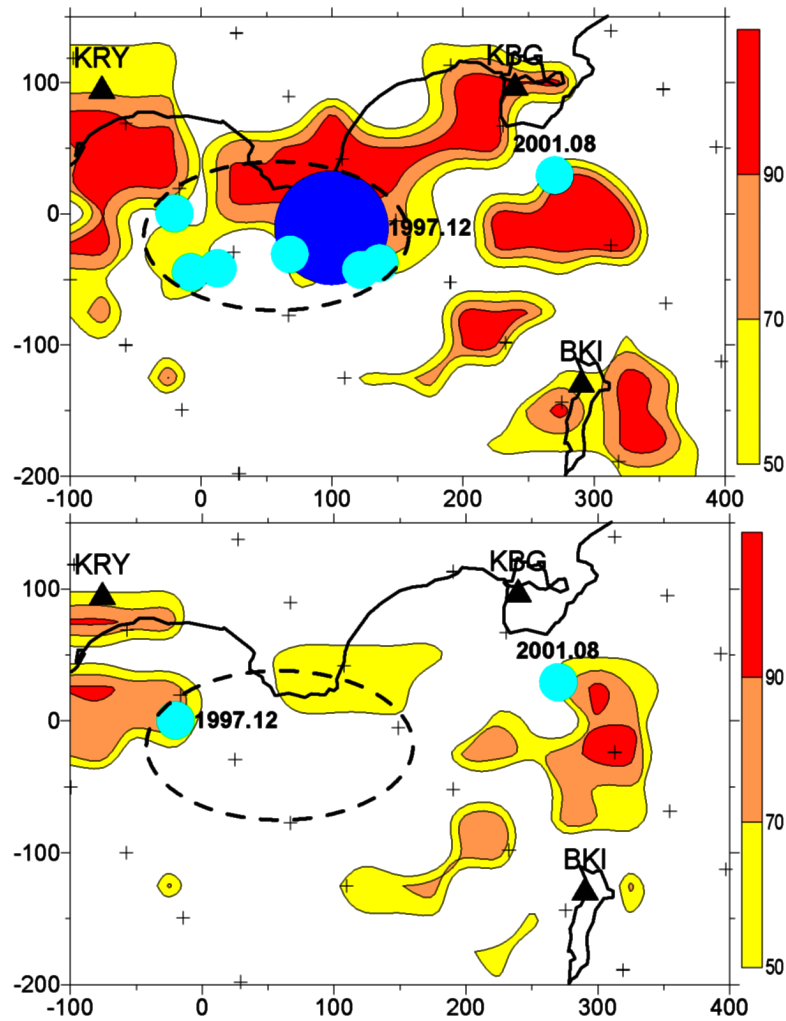


Figure 6. Fragments of EE maps of the Kronotski earthquake preparation area. The ellipse delineates the area of the strongest aftershocks recorded during three days after the main shock. The prognostic period begins in each map from January 1, 1997: (a) $K_{pr} \geq 13.5$, $H = 0-50$ km; (b) $K_{pr} \geq 13.5$, $H = 25-75$ km. The maps show epicenters of $K \geq 13.5$ earthquakes that occurred during the map serviceability periods. The rest of notation is the same as in Figure 3.

In the last decade, the feasibility of earthquake prediction has been widely discussed [Deshcherevsky et al., 2003; Geller, 1997; Geller et al., 1997; Kagan, 1997; Leary, 1997; Sobolev, 1999; Wyss, 1997]. In this discussion, both authors denying or doubting the earthquake prediction feasibility [Geller, 1997; Geller et al., 1997; Kagan, 1997; Leary, 1997] and authors adhering to an opposite point of view [Sobolev, 1999; Wyss, 1997] invoke rather weighty arguments in favor of their ideas, up to a change of a research paradigm in the field of earthquake prediction [Deshcherevsky et al., 2003]. As regards the question of whether earthquakes can be predicted, the author of the present paper believes that this question can be answered in two possible ways. (i) Since our knowledge of the earthquake preparation process is still incomplete and inadequate, earthquakes cannot be predicted. One should wait until the time when this knowledge becomes complete and only then start predicting real earth-

quakes. (ii) Based on the presently existing knowledge of the earthquake preparation process and known precursors and using available prediction algorithms, one should try to predict earthquakes, gaining in this way new knowledge and accumulating the prediction experience. Undoubtedly, such predictions should be probabilistic and their confidence levels should be estimated.

Here, we arrive at the following conclusion: in order to accumulate experience within the framework of the Code of earthquake prediction ethics [Sobolev et al., 1994], we can, in addition to the study of individual precursors, carry out tentative systematic scientific prediction of strong earthquakes on a real time scale, using available algorithms and precursors [Predictions,; Sadovsky, 1986; Sobolev et al., 1990, 1991].

Thus, long-term testing of the MEE algorithm in various seismically active regions has shown that the average prognostic effectiveness of the algorithm J_{MEE} is 2.5 times

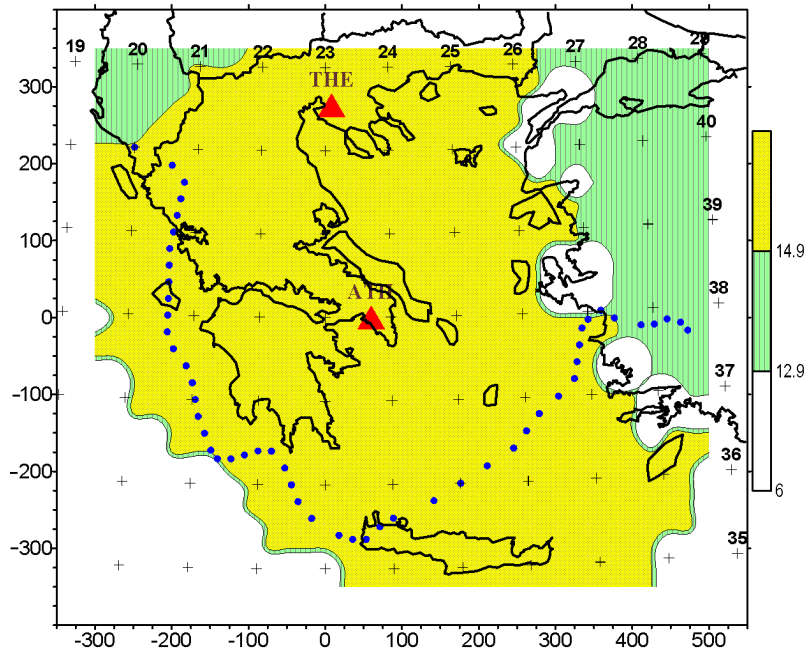


Figure 7. Map showing the distribution of the conditional probability of the $M_L \geq 5.5$ occurrence in Greece. The dotted line delineates the reliable recording area of $M_L \geq 3.5$ earthquakes. The legend shows shadings corresponding to various levels of the conditional probability $P(D_1|K')$ in percent. The rest of notation is the same as in Figure 3.

higher than in the case of random guessing. In this respect, we should note that about 68% of earthquakes involved in the prediction occurred in zones of the conditional probability $P(D_1|K) \geq 70\%$, and the average area of alarm zones amounted to 30% of the area where the seismicity level was one earthquake per year or more. These data are evidence of the real prognostic potential of the algorithm.

The most effective prognostic criterion used in the MEE algorithm is the density of seismogenic faults K_{sf} . Its prognostic effectiveness in five of nine tested seismically active regions was $J \geq 4$.

The case study of the Ust-Kamchatka, December 5, 1997 earthquake ($K = 15.5$) has demonstrated the MEE algorithm potentialities in relation to the source depth prediction of a future strong earthquake. The study in the Kamchatka region has also shown that the prognostic effectiveness of the MEE algorithm increases when the algorithm is applied to prediction of the strongest earthquakes.

The case study of western Turkey has shown that calculations with various sizes of averaging cells and comparison of their results with estimated sizes of the earthquake source and preparation areas as a function of the earthquake magnitude is helpful for more accurate adjustment of the MEE algorithm to prediction of earthquakes in a given range of magnitudes.

The long-term testing results indicate that the MEE algorithm can be recommended, on the one hand, as a means complementing observations of other precursors in inferred zones of high conditional probabilities (more than 70%) and, on the other hand, for taking preventive measures to miti-

gate the possible economic and social damage of a future strong earthquake.

6. Application of the MEE Algorithm to Prediction of Future Strong Earthquakes: A Case Study of Greece

The seismological database used for calculations was derived from the catalog of Greece earthquakes compiled at the geophysical laboratory of the Thessalonike Aristotle University. The catalog includes more than 40,000 earthquakes of local magnitudes $M_L \geq 1.5$ for the period from 1964 through 1993 and is representative (the contour $M_{rep} = 3.5$ encompasses most of the territory studied (Figure 7) [Smirnov, 1997]); these characteristics enabled the application of the prognostic MEE algorithm [Sobolev et al., 1990, 1991; Zavyalov et al., 1995].

Over the observation time from 1964 through 1992, 36 single earthquakes with $M_L \geq 5.5$ and their groups occurred in the region studied. The unconditional probability of the strong earthquake occurrence in this case was found to be equal to $P(D_1) = 0.1291$. To differentiate the study region in accordance with the unconditional probability level $P(D_1)$, we invoked data on the presence of seismically active faults in scanning spatial cells [Seismotectonic Map ..., 1989]. Using these data, we obtained the stationary conditional probabilities of the occurrence of a strong earthquake and false

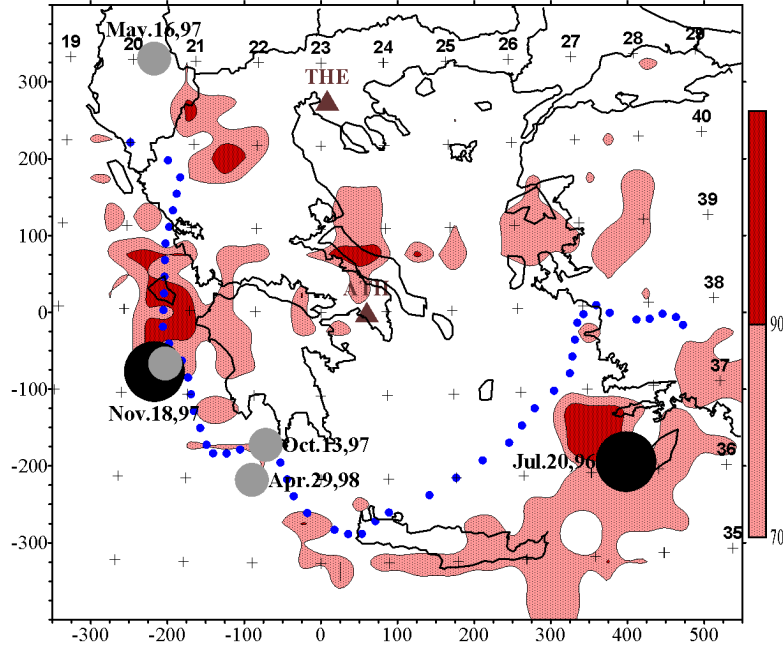


Figure 8. The Greece EE map for the period 1996–2002 calculated from earthquake catalog data over 1964–1995. The map shows the epicenters of strong ($M_L \geq 5.5$) earthquakes of 1996–2002 (five events). The rest of notation is the same as in Figure 3 and Figure 7.

alarm equal to $P(K'_i|D_1) = 0.8857$ and $P(K'_i|D_2) = 0.7288$, respectively. Using the inferred values of $P(D_1)$, $P(K'_i|D_1)$ and $P(K'_i|D_2)$ and the Bayesian formula (8), we calculated a map showing the distribution of the stationary conditional probability of strong earthquake occurrence over the territory studied (Figure 7). This map shows that the region is divided into three zones. The probability increases to 15% in cells including seismically active faults and decreased to 6% (about two times lower than the value $P(D_1)$) in cells in which no active faults are present. The probability remains the same (13%) in cells in which no data on seismically active faults are available.

Retrospective analysis of the whole series of Greece EE maps obtained at a step of one year for the period from 1978 through 1996 (the interval from 1964 through 1977 was used for training purposes) has shown that 48% of earthquakes in the prediction magnitude range $M_L \geq 5.5$ occurred in zones where the conditional probability level of such earthquakes exceeds 70%. In this case, the average alarm area with a seismicity level of $N_{|M_L \geq 3.5} \geq 1$ earthquake per year amounted to about 26% of the observation area.

Figure 8 shows the EE map with a prognostic period of 1996–2002 that was presented at the IASPEI General Assembly [Sobolev et al., 1997, 1999]. One of the main results obtained by these authors was high potential seismic hazard predicted for nearest years in areas east of the Athens-Thessalonica line (area A) and southwest of Athens (area B).

The prognostic period of the map terminated at the end of 2002. Analysis of strong ($M_L \geq 5.5$) earthquakes of the prognostic period in comparison with the EE map (Figure 8)

showed that four of five earthquakes of $M_L \geq 5.5$ (80%) occurred in zones of the conditional probability $P(D_1|K) \geq 70\%$. Two of these events (40% of the total number), the strongest ($M_L = 6.1$) earthquakes of July 20, 1996, and November 18, 1997, occurred in zones of $P(D_1|K) \geq 90\%$. The earthquake of November 11, 1997 occurred in the zone specified as potentially hazardous in August 1997 and this case can be regarded as a successful current prediction of a future earthquake [Sobolev et al., 1997, 1999]. The total area of all alarm zones in the map of Figure 8 amounted to $S_{al} = 45.3\%$ of the overall area with a seismicity level of 1 earthquake per year at $P(D_1|K) = 70\%$ and $S_{al} = 4.9\%$ at $P(D_1|K) = 90\%$. Therefore, the prognostic effectiveness of the Greece EE map of 1996–2002 in relation to strong earthquakes in zones characterized by a given level of $P(D_1|K)$ was $J_{MEE} = 1.77$ for $P(D_1|K) = 70\%$ and $J_{MEE} = 8.16$ for $P(D_1|K) = 90\%$. Note that the effectiveness of the EE map in the first case virtually coincides with the value obtained retrospectively from a long-term series of maps and is two times higher than this value in the second case (see Figure 4) [Zavyalov, 2002].

In the case of moderate earthquakes with $5.0 \leq M_L < 5.5$, only 12 of 23 events (52%) occurred in zones with $P(D_1|K) \geq 70\%$ and the effectiveness of the MEE algorithm at $P(D_1|K) \geq 70\%$ in relation to earthquakes of $M_L \geq 5.0$ was $J_{MEE} = 1.26$, which is somewhat better than the effectiveness in the case of random guessing. Therefore, it is apparently inappropriate to use the MEE algorithm for prediction of low magnitude magnitudes.

To sum up, we should note the following. Analysis of results of seismicity monitoring in Greece and adjacent areas

over 1996–2002 has shown that the prognostic MEE algorithm is applicable to the real-time medium-term prediction of strong earthquakes. The effectiveness of the MEE algorithm in relation to prediction of $M_L \geq 5.5$ earthquakes is no worse than its retrospective values (see Figure 4) and amounts to $J_{\text{MEE}} = 1.77$ and $J_{\text{MEE}} = 8.16$ at conditional probabilities of strong earthquake occurrence of 70% and 90%, respectively. Application of the algorithm to prediction of weaker earthquakes is less effective.

7. Further Improvement of the MEE Method

The MEE method described in this paper can be further improved and can be complemented with new physically and statistically substantiated precursors that meet the requirements described above and can be identified and used in prognostic practice with formal procedures developed for these purposes. In the opinion of the author, this approach offers a way for development and improvement of the method.

In particular, the following prognostic criteria can be included in a duly modified MEE algorithm: localization of the seismic process [Sobolev and Zavyalov, 1984b; Zavyalov and Nikitin, 1999, 2000], the traveltime ratio of P and S waves (parameter τ) [Slavina, 1976; Slavina and Myachkin, 1991; Sobolev and Slavina, 1977], the parameter K_{sf} including the fractal correction [Smirnov and Zavyalov, 1996], the earthquake grouping (clustering) parameter [Vasilyev, 1994; Zavyalov et al., 1995], and the RTL parameter [Sobolev and Tyupkin, 1996, 1997; Sobolev et al., 1996]. Any of these criteria meets the requirements formulated in the paragraph “Choice of Precursors.” For example, no procedures have been developed for mapping the seismicity localization parameter and estimating its retrospective statistical characteristics. Several precursors and prognostic algorithms (such as RTL) have recently appeared and procedures for their use in the MEE method have not been developed. However, these precursors are physically substantiated and, formal procedures of their identification being available, can be used in the MEE method.

8. Conclusion

This paper summarizes author’s investigations in various aspects of physics of the source and earthquake prediction in relation to the development of medium-term prediction method using a complex of physically substantiated prognostic criteria. Main results obtained in these studies are as follows.

Earthquake preparation models describing the behavior of precursors at various stages of the preparation process provided a physical basis and imposed several requirements to the choice of precursors for their use in the medium-term earthquake prediction method involving a complex of prog-

nostic criteria (the MEE algorithm). This paper addressed the following criteria that met these requirements and fracture this reason were accepted for their use in the MEE algorithm: the density of seismogenic faults K_{sf} , the recurrence plot slope γ , the number of earthquakes per unit time, and the released seismic energy $\sum E^{2/3}$. Their physical meaning and their relation to the earthquake preparation process are elucidated. Based on physical ideas and the experience of retrospective prediction using the aforementioned prognostic criteria, we developed a model describing the behavior of these parameters during earthquake preparation and proposed formal algorithms for their application in prognostic practice.

This paper proposed methods for collective treatment of precursors of various types (dynamic and quasi-stationary) are based on the spatio-temporal scanning and the Bayesian approach to the determination of the conditional probability of strong earthquake occurrence if values of several prognostic criteria are available, determination of their retrospective statistical characteristics, and estimation of the prognostic effectiveness for each precursor and for the algorithm as a whole.

Results of long-term testing of the MEE algorithm in various seismically active regions showed that its average prognostic effectiveness J_{MEE} is 2.5 times higher than in the case of random guessing. In this respect we should note that, on average, 68% of earthquakes involved in prediction occurred in the serviceability periods of expected earthquakes maps in zones where the conditional probability is $P(D_1|K) \geq 70\%$, and the average area of alarm zones amounted to 30% of the total area characterized by a seismicity level of no less than 1 earthquake per year. These results provide an idea of the real prognostic potential of the algorithm.

The estimates obtained in this study for the effectiveness of the precursors in use showed that the most effective prognostic criterion used in the MEE algorithm is the density of seismogenic faults K_{sf} . Its prognostic effectiveness was estimated at $J \geq 4$ in five of nine seismically active regions involved in the testing procedure.

Using various examples, the paper demonstrated the following aspects of the MEE algorithm applications:

- (i) it is applicable to prediction of the source depth of a future earthquake;
- (ii) it is particularly effective for prediction of the strongest earthquakes;
- (iii) the algorithm admits finer adjustment to prediction of earthquakes in a given range of magnitudes by comparing results obtained with various sizes of spatial averaging cells with size estimates of the source and earthquake preparation area as a function of the earthquake magnitude;
- (iv) the algorithm can be used for medium-term prediction of strong earthquakes on a real time scale.

Overall, this paper presented an algorithm for medium-term prediction of strong earthquakes using a complex of physically significant prognostic criteria, with prediction being characterized in terms of probabilistic estimates; the effectiveness of this algorithm is several times higher than that of random guessing prediction. Results of long-term testing indicate that the MEE algorithm is useful for iden-

tifying zones of a high (more than 70%) conditional probability in which observations of shorter-term precursors of other geophysical origins should be intensified and for taking appropriate preventive measures to mitigate economic and social damage that can be caused by a future strong earthquake.

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