Relaxation creep model of impending earthquake

Vitali A. Morgounov
Institute of Physics of the Earth, Russian Academy of Sciences, Moscow, Russia

Abstract
The alternative view of the current status and perspective of seismic prediction studies is discussed. In the problem of ascertainment of the uncertainty relation Cognoscibility-Unpredictability of Earthquakes, priorities of works on short-term earthquake prediction are defined due to the advantage that the final stage of nucleation of earthquake is characterized by a substantial activation of the process while its strain rate increases by the orders of magnitude and considerably increased signal-to-noise ratio. Based on the creep phenomenon under stress relaxation conditions, a model is proposed to explain different images of precursors of impending tectonic earthquakes. The onset of tertiary creep appears to correspond to the onset of instability and inevitably fails unless it is unloaded. At this stage, the process acquires the self-regulating character and to the greatest extent the property of irreversibility, one of the important components of prediction reliability. Data *in situ* suggest a principal possibility to diagnose the process of preparation by ground measurements of strain-rate-dependent parameters, like electromagnetic emission, etc. Laboratory tests of the measurements of acoustic and electromagnetic emission in the rocks under constant strain in the condition of self-relaxed stress until the moment of fracture are discussed in context. It was obtained that electromagnetic emission precedes but does not accompany the phase of macrocrack development.

Key words *earthquake – focal zone – creep of rocks – relaxation of load – short-term precursors*

1. Introduction

Discrepant opinions on earthquake predictability (Evance *et al.*, 1996; Geller, 1997) reflect the state of fundamental investigations into rock fracture process. In fact, it is difficult to define accurately the term «fracture». Each of the present theories of strength describes a particular type of material behavior at a certain type of load, whereas a real rock mass is in a state of elastic compression, plastic flow, and brittle failure simultaneously.

Mailing address: Dr. Vitali A. Morgounov, Institute of Physics of the Earth, Russian Academy of Sciences, Bol'shaya Gruzinskaya 10, Moscow 123810, Russia; e-mail: vam@uipe-ras.scgis.ru

Moreover, elasticity, plasticity, and brittleness characterize not only a material but also the conditions of its deformation. Thus, discussions on the chaotic and highly nonlinear nature of the nucleation process, *i.e.* self-organized criticality that has no length; high heterogeneity of the medium; uncertainty in spatial distribution of precursors and etc. seem to be reasonable.

The difficulty lies not only in the inhomogeneity of the medium as in the uncertainty in the conditions, at which a deformed volume is loaded, and in rheological properties of rocks and their variations during the deformation of the inclusion and host medium, which exhibit to a varying extent, elements of elasticity, plasticity, and brittleness, depending on time and space.

Geophysical scales of the process at hand and near-critical stresses existing in the Earth's crust and near the surface make it necessary to account the real rheology of geomaterials, which, as is well known, is characterized by clearly pronounced inelastic properties. The influence of steady tectonic forces, overburden pressure, temperature is favored by the development of creep in rocks.

Similar to the qualitative difference between the process of accumulation of the elastic energy in the specimen at the initial stage of deformation under constant load and the release of this energy in the form of inelastic strain (creep) before its failure in the case of earthquake, we ascertain the same succession for long-intermediate precursors (accumulation of elastic energy) and short-term precursors indicating the release of energy immediately before the shock. Thus, a comparison suggests itself between tertiary creep and short-term precursors.

It is known that the time scale of the final, tertiary stage of creep is characterized by an avalanche-like increase in the strain rate immediately before the failure. This stage is completely nonlinear; however it is not chaotic. The process in the focal zone during this stage acquires, to a large extent, a self-regulating character.

As the strain rate in the focal zone increases by orders of magnitude one can expect the strain process would envelop at some extent the surrounding volume (sensitive or preparation zone) that would increase the signal/noise ratio of the measuring parameters on the Earth's surface. Therefore, the impending failure is easier to detect at this stage by monitoring a complex of short-lived geophysical fields, *e.g.*, electromagnetic (EME), acoustic (AE) emission and etc. The process at the final stage acquires to the maximum extent the property of irreversibility and avalanche-like growth, the most important component of the prediction problem. Practical significance and methodological advantages of the identification of the final stage of the earthquake preparation process determine the priorities of the short-term prediction.

Different approaches have been made to model the earthquake nucleation process. Brady, (1976) and Rice and Rudnicki (1979) laid down the foundations of the «inclusion model» according to which the zone of strain weakening material is embedded in nominally elastic surroundings subjected to tectonic stress. Thus, the whole deforming volume surrounded the future hypocentre can be divided into two basic parts: focal zone where inelastic strain is developing (creep), and the outer one, subjected to quasi-elastic deformation (preparation zone). We use the term «quasi-elastic embedding medium» to indicate the capability of rocks to transmit the orders of magnitude of elastic strain through the earthquake preparation area and at the same time a manifestation of the inelastic component and therefore the appearance of the EME precursors (Gokhberg *et al.*, 1982; Yoshino *et al.*, 1992; Morgounov, 1993a,b, 1996; Morgounov *et al.*, 1994; Hayakawa and Fujinawa, 1994; Varotsos *et al.*, 1996; Vallianatos, 1998; Vallianatos and Tzanis, 1998, etc.) and AE (Morgounov *et al.*, 1991) etc.

The present paper entails the objectives which concern the modeling of the strain process in the focus of the impending earthquake. Based on the analyses of the tertiary creep under relaxation load, a model of the final stage of earthquake nucleation is proposed. The relaxation stress mechanism during creep of rocks of the inclusion results in the delay of the failure instant, *i.e.* to prolong the precursory duration and be the cause of the variety of the precursory images. The calculations are compared with the data *in situ*. The results of the laboratory tests of rocks under constant strain in the condition of self-relaxed stress are discussed in the context of the model.

2. Tertiary creep in rocks and prediction of the fracture

2.1. *Creep under load relaxation conditions*

Creep of rocks, developing in laboratory conditions as well as in crustal rocks *in situ* on a geological time scale, are divided into three main stages: attenuation, steady, and accelerated (or tertiary), with a characteristic, avalanche-like increase in the rate immediately before failure.

Time becomes a very important parameter at the final nonlinear stage of the earthquake nucleation process. It is well known that the strength of most brittle materials, including rocks, is time dependent. The weakening in time is known as static fatigue. For a wide variety of materials, static fatigue is due to stress corrosion (Scholz, 1968). Griggs (1940) investigated creep of alabaster and came to the conclusion that it is possible to have flow cleavage and fracture cleavage in the same region by developing flow cleavage in the layers, where the amount of deformation is below the critical value for fracture, and fracture cleavage, where the deformation exceeds the critical value. He concluded that the application of the laboratory experiments may be made to rock fracture in earthquakes, in particular for the determination of the final stage – creep before rupture – the period of increasing strain rate.

Benioff (1951) pointed out that the consideration of many seismic phenomena is based on a simple elastic theory, in which strain is proportional to stress and is independent of time. However, the characteristics of rocks depart greatly from these simple assumptions. Benioff proposed the creep theory of aftershocks and presented evidence that aftershocks are produced by creep of fault rocks.

Kranz and Scholz (1977) emphasized that the third stage, (tertiary creep) has received comparatively little attention. They defined tertiary creep as that part of creep behavior, in which the strain rate is continuously increasing. Once a rock has entered this stage it inevitably fails unless it is unloaded. Thus in a sense the onset of tertiary creep appears to correspond to the onset of the instability that leads to fracturing. Their hypothesis of the critical crack density suggests that the onset of tertiary creep will occur when a critical value of inelastic volumetric strain is achieved. Wu and Thomsen (1975), Lockner and Byerlee (1980) demonstrated a correlation between acoustic emission and the onset of tertiary creep and have shown that the onset is marked by supraexponential rise in the activity of acoustic emission, generally coinciding with an increase in strain rate.

At its final, avalanche-like stage, the process acquires a self-regulating character. Unlike the term «self-organized criticality», the term «self-regulating creep» assumes a particular meaning and testifies the basic possibility to diagnose the nucleation process.

Experimental data show that a seismic event can be preceded, and in some cases accompanied by a relatively moderate value of the pre-seismic anomalous signal. In a number of cases, the moment of the earthquake is not accompanied by any specific disturbances or all the more by a burst of the signal. (At least co-seismic anomaly is of the order of magnitude of pre-seismic one). Moreover, many seismic events occur during a drop phase in a bay-shaped pre-seismic anomaly or after the anomaly has terminated (so-called «quiescent» phase of the signal).

A precursor signal recorded at the Earth's surface is an indirect response of the upper crust to deformation processes at the focus. Short-lived geophysical phenomena such as pulsed electromagnetic emission (EME), acoustic emission (AE), electrical resistivity (ρ), hydrogeochemical parameters, etc., are controlled by strain rates in the vicinity of an observation point. This assumption, together with experimental evidence, leads to the conclusion that in some cases the final stage of the nucleation process in the focal area is not of the avalanche-like type.

In such cases, the classical model of creep, *i.e.* a progressive deformation under constant load with continuously increasing strain rate, as is the case of laboratory conditions, cannot serve as a direct analog of the source processes. The difference between the uniaxial loading of a specimen in the laboratory and the focal inclusion of rock *in situ* is that the load in the natural condition does not remain constant during progressive inelastic deformation of the inclusion due to the redistribution of the external tension applied to the inclusion among the host rock environment, *i.e.* the load decreases in the process of deformation of the inclusion.

Laboratory tests (Glushko and Vinogradov, 1982; Kranz and Scholz, 1977) indicate that the time of accelerated creep in rock samples (argillite, granite and quartzite) does not exceed 10-30% of the total time of deformation (~ 8-10 h) under a load amounting to 80-95% of the failure value. Similar values were obtained in analyses of the immediate-term precursors *in situ* (Morgounov, 1993a,b, 1996).

Thus, at the third stage of creep, rocks of an inclusion reach maximum rates of plastic deformation during a relatively short period of time. On the one hand, this can result in a drop of elastic tectonic stresses and, on the other, an increase in counteractive stresses in host rocks, which takes up the load in the process of fast deformations of an inclusion.

If F_1 is the value of elastic tectonic forces that decrease as plastic deformation in the inclusion volume V increases and if F_2 stands for forces of host rock counteraction, which increase with deformation of the inclusion, the resulting force $F(\varepsilon) = F_1 - F_2 \neq$ const decreases with the increasing strain $\varepsilon\,(t)$. Taking into consideration the elastic nature of tectonic stresses, we represent $F(\varepsilon)$ in the form of a linearly diminishing function

$$F(\varepsilon) = F_0(1 - \varepsilon(t)/\varepsilon_x), \tag{2.1}$$

where F_0 is the initial load at the onset of accelerated creep and, ε_x is the strain at which elastic forces relax, *i.e.* $F(\varepsilon_x) = 0$.

Under natural conditions, the redistribution of the stress and strain between the host medium and an inclusion is complicated and nonstationary, and gives rise to the diversity of precursors observed. Nevertheless, its main phases are recognizable in terms of a simplified model.

The empirical dependencies of strain $\varepsilon\,(t)$ for the steady-state and accelerated creep stages are expressed through power and exponential functions (Kennedy, 1962)

$$\varepsilon(t) = \varepsilon_0 + \varepsilon_\Lambda t^n + \varepsilon_1 \exp\,(\alpha t), \tag{2.2}$$

where ε_0 – initial strain, ε_Λ, ε_1 – scale coefficients.

Comprehensive studies of structural materials have established that the stage of accelerated creep ($n > 1$) consists of two, substantially differing phases (Kennedy, 1962) and this was confirmed by experiments on rocks (Ohnaka, 1983). At the beginning of the accelerated stage, the deformation process, as well as steady-state creep, is described by the power law. In the course of time, the process is accelerated and, during the time interval directly preceding the brittle failure, a law close to the exponential one could govern it.

The assertion that, unlike long- and intermediate-term precursors, short-term ones are controlled by the release of previously accumulated energy is not questioned. Moreover, the fact that short-term precursors are caused by inelastic crustal deformation is equally unquestionable. Nevertheless, the models based on the principles of static theory of elasticity, in which $\varepsilon \sim \sigma$ are used for the analysis of processes of this scale, even though this is not correct even as a first approximation.

A general relationship between creep rate and stress renders possible the transition from a region of small stresses to one of high stresses (Kennedy, 1962)

$$\varepsilon'(t) = B\sigma^m + b\exp(\beta\sigma) \tag{2.3}$$

where ε' is the strain rate; $\sigma = F/S$ is the stress in a cross section S of the volume at a load F, the values of $m \geq 1$ and B, b, β are constants.

The problem concerning the estimation of the stress state $\sigma(t)$ in a cross-section of an impending fracture is complicated and has not been solved in the general form even for the simplest case of necking in a specimen of structural material produced by extension. Laboratory tests indicate that a decrease in the carrying capacity of specimens is due to the shortening of a cross-section that actively supports a load in the process of progressive fracture; rock deformation characteristics remain more or less constant. The moduli of elasticity and deformation, determined from the σ-ε curves, characterise not the properties of material but the rigidity of a system into which specimens deformed beyond the strength limit are changed.

2.2. *Fracture as an act of breakage of internal bonds*

According to Benioff (1951), creep may be purely compressional, purely shearing, or a combination of the two. In the latter case, the compressional phase occurs first. The shearing phase follows after an interval of 0.01 to 2.4 days.

As noted by Tapponnier and Brace (1976), almost all cracks are tensile in nature. Small amounts of shear were seen to occur only where cracks coalesce. This may be because the shear stress is enhanced in the region between approaching crack tips. Kranz (1979) confirmed these results.

External forces, classified as compressional or shear, result to the maximum tensile stresses in the material, along the direction of which internal bonds are broken once critical values are attained. Outcrops of a rock mass subjected to multiaxial compression are shown to exhibit fracture patterns consistent with the stress state of general extension. It seems that in rock mechanics, regardless of the level of observations, the term «fracture» defines such a phenomenon, as the failure of internal bonds.

Therefore, fracturing is the process of a gradual accumulation of defects (cracks) and a decrease in the carrying capability. In the case of extension this means, rather than a decrease in the strength of the material, a decrease in the cross-sectional area of a body that actually sustains the applied loads. In essence, this process is connected with the amount of broken internal bonds in the fracture zone, *i.e.*, the surface on which a rupture will occur. Because of the lack of exact solutions to this problem, we will consider the phenomenological approach.

The term «effective cross-section» S_{eff}, will be understood as a diminishing surface across which, due to the weakening of internal bonds, a fracture eventually propagates. Then

$$\sigma = F/S_{\mathrm{eff}}. \tag{2.4}$$

The S_{eff} values are controlled by the degree of the injury of carrying capacity of the material in the cross-section of future rupture, rather than by the visible geometric dimensions of a cross-section. This notion does not correspond to the term «necking» since some materials fail without necking even during extension. The rupture occurs on a surface on which cohesive forces control the dynamics of the fracture. Because shearing is equivalent to a combination of compression and extension, in case of shear (and compression), the definition of S_{eff} does not change. A S_{eff} reduction with time is associated with the injury of the material, an increase in stress σ, and an acceleration of the fracturing process in the case of constant load.

2.3. *Short-term earthquake precursors*

Formulae (2.2) and (2.3) permit the consideration of two phases of accelerated creep, observed experimentally. The initial phase of the accelerated stage, responsible for short-term precursors can be described by generally used power functions (n, $m > 1$)

$$\varepsilon(t) = \varepsilon_0 + \varepsilon_\Delta t^n; \qquad \varepsilon' = B\sigma^m. \tag{2.5}$$

At the load F_0 = const, we have

$$S^m_{\mathrm{eff}} = BF^m_0 n^{-1}(\varepsilon\text{-}\varepsilon_0)^{(1/n-1)}(\varepsilon_\Delta)^{-1/n}. \tag{2.6}$$

With an increase in plastic deformation of an inclusion V, the elastic tectonic load decreases. A change in the deformation pattern entails, in general, a variation in the physical parameters of the deformed volume, including $S_{\mathrm{eff}}(\varepsilon)$. In order to clarify the general features, we assume that the $S_{\mathrm{eff}}(\varepsilon)$ dependencies are the same for dropping and constant load. This is not quite correct, but does not change the essence of the statements to be illustrated. (According to Kranz, 1979, there can be a difference in the mode of crack development between tests at constant stress and at constant strain rate).

To determine the conditions of fracturing, we will use the second theory of strength, in terms of which fracturing occurs when the ultimate values of tensile and compression strain are reached: $\varepsilon_{\mathrm{tens}} \leq \varepsilon_f \leq \varepsilon_{\mathrm{compr}}$. The statement is consistent with the results of rock tests, which indicate that inelastic volumetric strain at the onset of tertiary creep is found to be independent of the stress level (Kranz and Scholz, 1977).

Taking into account (2.1), (2.5), and (2.6), we obtain

$$\varepsilon' = n(\varepsilon_\Delta)^{1/n}(\varepsilon\text{-}\varepsilon_0)^{(1-1/n)}(1-\varepsilon/\varepsilon_x)^m. \tag{2.7}$$

The plots of $\varepsilon(t)$ and $\varepsilon'(t)$, presented in fig. 1 for F_0 = const and $F(\varepsilon) = F_0(1-\varepsilon(t)/\varepsilon_x)$, show that under conditions of dropping load at the initial stage, the strain $\varepsilon(t)$ grows according to a power law, reaches a point of inflection and asymptotically tends to a constant value. The strain rate curve $\varepsilon'(t)$ reaches its maximum at the point of inflection of the curve of the strain and then tends to zero. Depending on the rheological properties of geomaterial and on the loading mode, the critical strain value is attained at different points of the $\varepsilon(t)$ curve, which determines the resulting shape of a precursor.

The effect of the «earthquake during the drop of an anomaly» thereby receives its natural explanation. Indeed, as short-lived fields like EME, AE, etc. are controlled by strain rate, in the condition of relaxing load the intensity of these parameters following the curve of strain rate (fig. 1, right dashed line), after the passage of its maximum, *i.e.* the inflection point of the strain curve (fig. 1, right solid line), goes down to the moment when the strain reaches the value of the strength limit ε_f (fig. 1), *i.e.* till the moment of the earthquake.

The diagram in fig. 2a,b illustrates the conditions responsible for various shapes of the anomaly. The curves of fig. 2a strain ε and fig. 2b strain rate ε' correspond to the shapes of observed geophysical anomalies at different loading modes: 1 = constant load F_0; 2 = curve with a linear interval of ε; 3 and 4 = relaxing loads of different duration, and 5 = anomaly not associated with an earthquake (false alarm).

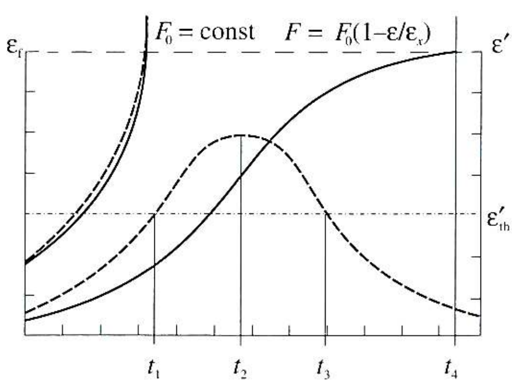

Fig. 1. Schematic view of the curves of strain ε (solid lines) and strain rate ε' (broken lines) at a constant load F_0 and at a relaxing load $F = F_0(1-\varepsilon/\varepsilon_x)$. The dot-and-dash line indicates the threshold level (ε'_{th}) of strain rates accessible to geophysical measurements. The strain ε_f is the fracture (rupture) threshold. The times t_1, t_2, and t_3 are those of the beginning, maximum, and termination of the anomaly, and $t_3 - t_1$ is its duration. t_4 indicates the moment of the earthquake.

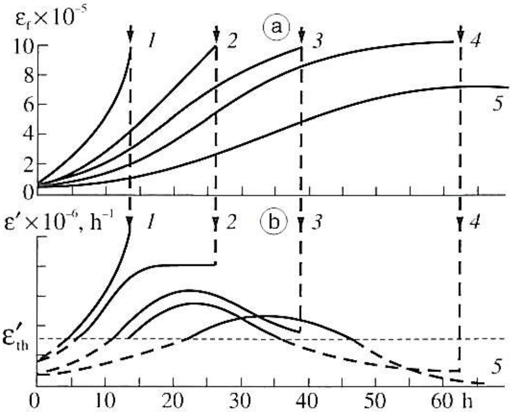

Fig. 2a,b. Illustrative curves of (a) strain ε and (b) strain rate ε', *versus* time corresponding to the shapes of recorded geophysical anomalies for various modes of loading: 1 = constant load F_0; 2 = curve with a linearity interval of ε; 3, 4 = relaxing load of different duration; 5 = anomaly unrelated to an earthquake (false alarm).

Because the stage of accelerated creep has been poorly studied, only qualitative estimates can be made. We will illustrate the approach in question for the values of $m = 2$, $n \sim 1$. In this case, we obtain from (2.7) the formula for estimating the duration of creep under conditions of relaxing load

$$\Delta T = \varepsilon_\Delta^{-1}\varepsilon_x^2(\varepsilon_f - \varepsilon_0)(\varepsilon_x - \varepsilon_f)^{-1}(\varepsilon_x - \varepsilon_0)^{-1} \quad (2.8)$$

where ΔT is the time interval between the onset of accelerated creep ε_0 and the moment of fracture ε_f.

In order to estimate the deceleration rate of the deformation process due to the load relaxation we use data on the accelerated creep stage, obtained *in situ* and by laboratory experiments. According, for example, to Glushko and Vinogradov (1982), at a constant load amounting to ~ 80% of the failure value, the third stage of creep in argillite lasts for τ_a ~ 8-10 h. The preliminary estimation of the minimal duration of tertiary creep *in situ* yields a similar value ~ 10 h (Morgounov, 1996). At $\varepsilon_0 = 0.1\,\varepsilon_f$, $\varepsilon_\Delta \sim (\varepsilon_f - \varepsilon_0)/\tau_a h^{-1}$ and $\varepsilon_x = 1.1\,\varepsilon_f$ (10% of the load relaxation strain), we obtain $\Delta T \sim 120$ h, which corresponds to the duration of recorded anomalies (Corwin and Morrison, 1977, etc.) and to the time scale of the short-term precursors of the bay type.

These estimates of ΔT, which do not require absolute values of ε_0, ε_x and ε_f, depend strongly on the ratio $\varepsilon_x/\varepsilon_f$. At $\varepsilon_x/\varepsilon_f \gg 1$ the effect of an increase in the duration of deformation caused by stress relaxation vanishes. At $\varepsilon_f \rightarrow \varepsilon_x$, ΔT rapidly increases. In particular, at $\varepsilon_x = 1.05\ \varepsilon_f$ (5% of the load relaxation strain), $\Delta T \sim 230$ h. The stress drop results in a decrease of the strain rate at the source, deformation becomes more unstable and can stop. The anomalies at this period may be classified as false alarms. This may be considered as a gradual transition to immediate-term precursors, since the remaining strain $\Delta\varepsilon = \varepsilon_f - \varepsilon(t)$ is realized at a shorter time.

This mechanism of alternating activation and relaxation of the creep process can explain precursors of the oscillatory type with varying periods, often observed in experiments. The instability of the fracture process with respect to input parameters could explain a scatter in the parameters of precursor occurrence. Particularly, in the case of the value ε_x is close to the value of strength limit ε_f, *i.e.* $\varepsilon_f/\varepsilon_x \rightarrow 1$, the load $F \rightarrow 0$, (2.1) and the process become more unstable due to the growing influence of the heterogeneity of the Earth crust. It could result in a higher instability of the nucleation process (scattering the precursor's parameters) for weaker earthquakes in comparison with more powerful ones.

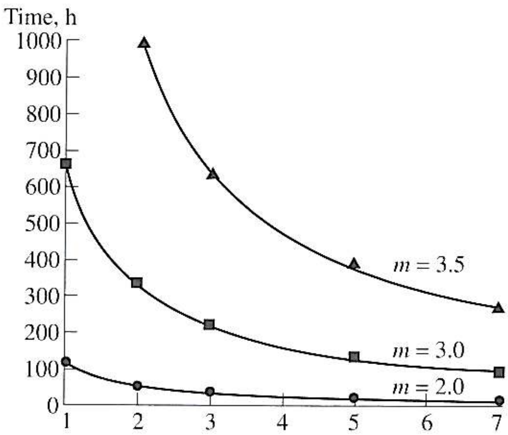

Fig. 3. The solution of eq. (2.7) at the accelerated stage of creep under a relaxing load, depending on the exponents n and m.

The solutions to eq. (2.7) for $\Delta T(m, n)$ with the exponents n = 1-7 and m = 2-3.5, characterizing average properties of material and loading conditions, are plotted in fig. 3. The duration of the deformation process under a load that relaxes with increasing m changes into time scale of intermediate term precursors, which have the same characteristic features and modes of occurrence as the short-term ones. On the other hand, the curves tend asymptotically (with increasing n and decreasing m) to the duration of creep at constant load, which indicates the existence of a minimum duration of the fracture process.

2.4. *Immediate precursors of an earthquake*

The initial phase of the accelerated creep stage in the crust may apparently be described by the above power-law (2.5) representation of $\varepsilon(t)$. Depending on the local conditions, loading mode, depth and size of the focal zone, the plastic properties of rocks can vary significantly. Under specific *in situ* conditions, the third stage of creep can reach an avalanche-like phase, which is described by the second terms in formulae (2.2) and (2.3). However, taking into account the scales of *in situ* deformation processes and the absence of bursts of the signal on the onset of earthquake (co-seismic signal is the order of pre-seismic), we use a less steep power relation for $\varepsilon'(\sigma)$, *i.e.*

$$\varepsilon(t) = \varepsilon_0 \exp(\alpha t), \qquad \varepsilon' = B\sigma^m. \qquad (2.9)$$

In view of (2.1), this yields

$$\varepsilon' = \alpha\varepsilon(1 - \varepsilon/\varepsilon_x)^m. \qquad (2.10)$$

This equation can be related to immediate precursors. Numerical assessments are complicated by an uncertainty in α and m. The solution of (2.10) for $m = 2$ yields an estimate for the time

ΔT of the terminal stage of creep under the condition of stress relaxation:

$$\Delta T = \alpha^{-1}[\varepsilon_x(\varepsilon_f - \varepsilon_0)(\varepsilon_x - \varepsilon_f)^{-1}(\varepsilon_x - \varepsilon_0)^{-1} + \quad (2.11)$$
$$+ \ln \varepsilon_f \varepsilon_0^{-1}(\varepsilon_x - \varepsilon_0)(\varepsilon_x - \varepsilon_f)^{-1}].$$

Considering that the ratio of duration of the avalanche-like stage to that of the accelerated stage is $\tau_{av}/\tau_{acc} \sim 0.5\text{-}0.2$ (Onhaka, 1983), we have $\alpha = \ln(\varepsilon_f/\varepsilon_0)/\tau_{av} = 0.46\text{-}1.1\ \mathrm{h}^{-1}$. For the previously assumed relations $\varepsilon_0 = 0.1\ \varepsilon_f$ and $\varepsilon_x = 1.1\ \varepsilon_f$, $\Delta T \sim 30\text{-}12$ h. Despite the uncertainty of initial data, the use of the exponential dependence $\varepsilon(t)$ leads to different time scales that are closer to the estimates for the immediate precursor time scale (Morgounov, 1996).

More elaborated estimation is difficult without additional information on the initial parameters. Even such a parameter as the fracture strain ε_f depends on rock type and its *in situ* and laboratory estimates differ by orders of magnitude, from 10^{-5} to 10^{-3}. However, this does not seem to be an insurmountable obstacle for prediction investigations, since relative values of the parameters in question are more important. The dependence of time intervals on α and m is illustrated by the curves plotted in fig. 4.

Based on the above ideas, we consider a typical example of anomalous signal records

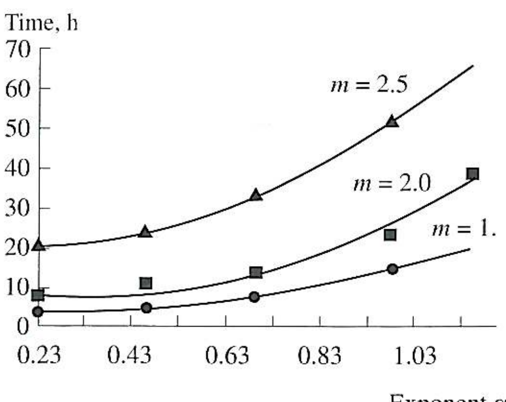

Fig. 4. The solution of eq. (2.10) at the avalanche-like stage of creep under a relaxing load, depending on the exponents α and m.

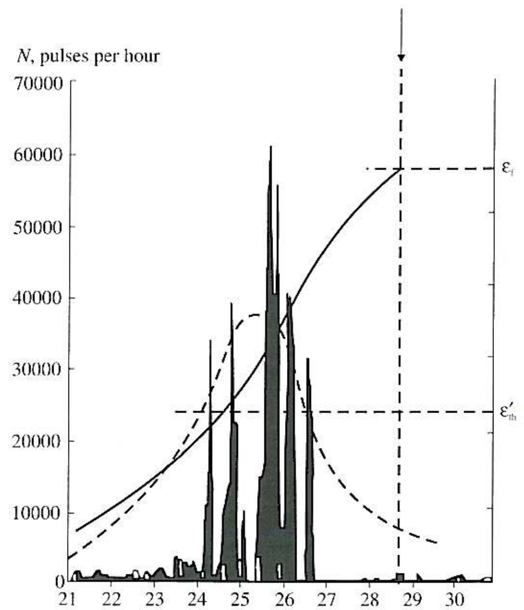

Fig. 5. An example of the EME anomaly interpretation: ε'_{th} is the threshold value of the measurable strain rate; ε_f = fracture strain; N = number of pulses per hour at the selected threshold of sensitivity. The down arrow indicates the moment of the November 28, 1992, $M = 4.0$, earthquake at the Mineralnye Vody, Caucasus, research site at a distance of 45 km from the point of measurement. The horizontal coordinate is represented in days. See explanations in the text.

of pulsed electromagnetic radiation (Zdorov, and Morgounov, 1997), which were employed for empirical earthquake predictions at the Mineralnye Vody, Caucasus, research site. The superposition of the theoretical curves $\varepsilon(t)$ and $\varepsilon'(t)$ onto the experimental records of the EME, shown in fig. 5, illustrates the interpretation of the pulsed EME anomaly in terms of relaxation creep model. Intensive EME was detected a few days before the earthquake that can be interpreted by the excess of strain rate $\varepsilon'(t)$ its threshold level ε'_{th} with the maximum near the inflection point of the strain curve $\varepsilon(t)$. The arrow indicates the moment when the strain attained the value of strength limit ε_f, *i.e.* the onset of the earthquake $M = 4.0$, occurred at ~ 45 km from the observation point.

In this case, the precursor duration was $t_3 - t_1 \sim 54$ h, with the time of anomaly detection before an earthquake being $t_4 - t_1 \sim$ 4.5 days. Here the notations of fig. 1 are used: t_1 – the onset of the anomaly ~ 06 h LT, November 24, 1992; t_3 – the end of the anomaly ~ 12 h LT, November 26, 1992; t_4 – the onset of the earthquake 15 h 28 min. LT, November 28. The parameters of the smoothed shape of the pre-seismic anomalous signal, the representative values of ε_0, ε_f, within the framework of the model proposed (2.7), (2.10) enable us to formulate the inverse problem for the determination of the inelastic properties of the rocks (m, n, α) in a particular seismic event.

The above examples do not exhaust the possible types of precursors, which reflect the specific conditions at the focal zone. In some cases, the described processes or its parts can repeat, thereby generating an oscillatory sequence of the precursor occurrence. The study of these features may be instrumental in improving the accuracy of the predicted parameters of earthquakes and other geodynamic events.

3. Laboratory modelling of relaxation creep in the rocks

Laboratory modeling of the earthquake nucleation process is fairly criticized because of the lack of correspondence between the time-spatial scale, strain rates in laboratory and *in situ* conditions. But the laboratory experiments are informative in the case when the basic relationships of the destruction process are studied. In laboratory tests of rock creep, the primary and secondary stages were studied. At the same time Griggs (1940), Kranz and Scholz (1977), Lockner and Byerlee (1980) successfully used the rock creep tests in the laboratory for modeling of earthquake nucleation.

The stage of tertiary creep is a specific question of the problem. In a sense, the onset of tertiary creep appears to correspond to the onset of the instability that leads to fracture. This is one of the most important components of the reliability of the prediction task. The critical crack density has been invoked to explain the initiation of macro fracturing. According to Kranz (1979), the stage determines the onset of tertiary creep, where crack coalescence becomes more important than the slow growth of individual cracks. Wu and Thomson (1975), Lockner and Byerlee (1980), Ohnaka (1983) defined the onset of tertiary creep by a law close to exponential rise of acoustic emission.

Unlike EME, the methods of acoustic emission are commonly used for the studies of the destruction mechanisms in laboratory conditions and for the purpose of the safety work a mine. For the purpose in question the combined usage of both methods is a subject of specific interest and we will discuss the results of the experiment of uniaxial compressing of argillite specimens under the condition of relaxing loading with the registration of AE and EME (Morgounov *et. al.*, 2001), fig. 6.

A closed-loop, servocontrolled hydraulic testing machine, Instron-1275, was used to study the EM and acoustic effects during the failure of cylindrical samples of argillite, 10 cm long and 5 cm in diameter. The AE and EME gages were placed on the sample. The EME gage represented 15 coils of wire 0.14 mm in diameter coiled around the sample (a gage of the electric type). As the AE gage a wide-range ultrasonic gage was applied. The transformation coefficient of the gages chosen was constant within the 1-100 kHz frequency band.

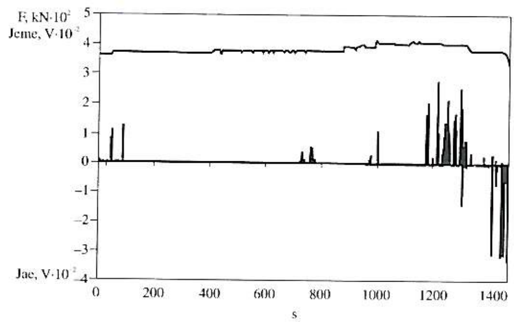

Fig. 6. Plots of axial load (upper curve), amplitudes of electromagnetic and acoustic (reversed axis) pulses *versus* time (in seconds) during the final stage of destruction of argillite specimen under in steps-constant strain regime of load. The peak values of amplitudes of Jae and Jeme are given in Volts $\cdot 10^{-2}$ reduced to the sensor's output. From $t \sim 1100$ s the initial strain remained constant.

Soon it was found that the 1-100 kHz frequency range is informative for investigation of the final stage of destruction. In particular, Warwick *et al.* (1982) reported that the appearance of the low-frequency component of the electric field with a maximum at a frequency of 50 kHz is associated with the destruction process. Ogawa *et al.* (1985) recorded the maximum intensity during the destruction of rock at the frequency of 30 kHz.

The signals from the guages of load of the movement of the active grip of AE and EME were input to ADC and, then, upon amplification, the signals were input to peak detectors. The frequency of passage of packets of signals and the maximum amplitude in a wave packet were recorded by the AE and EME channels with a timing interval of 0.5 s.

The loading of a sample was carried out stepwise with a lag at each step, while the loading trajectory was controlled by the deformation. When the axial deformation reached a certain value determined from preliminary tests, it was fixed in the sample, and the destruction process of the sample material was further observed in the process of relaxation of the load.

Such tests were chosen to slow down the destruction process and to make modelling of the loading condition *in situ* more accurate. Unlike the ordinary creep process under constant load, where the external energy is supplied to the sample while deforming the sample, in this case the energy accumulated initially in a sample decreases during the process of destruction and relaxation of external load and, consequently, the short final stage of the fracturing process is prolonged. This kind of loading was previously used by Peng (1973), who obtained in particular that the duration of creep rupture was inversely proportional to creep load.

For the purpose of testing the threshold, the instrument was adjusted for a higher value of input signal to register more powerful acoustic impulses inherent to the final stage of fracturing process. It was another reason to measure acoustic and EM emission in the lower frequency band of 1-100 kHz (Warwick *et al.*, 1982; Ogawa *et al.*, 1985).

Figure 6 shows a representative example of the acoustic (J_{AE}) and electromagnetic (J_{EME}) emission rates. The final 10 min of the test are the most important in the experiment. Up to the time of $t \sim 1100$ s, the specimen was subjected to a step deformation to the level $\sigma \sim 0.9$-0.95 of its strength and the strain 0.025. Beginning with the moment of the step loading the process of destruction of the specimen was going under a fixed deformation (fixed positions of grips) in the regime of self-organized stress relaxation.

From fig. 6 we can conclude, that from $t \sim 1100$ s, when the initial strain was constant, consistently without adding external energy, the process of failure developed with opening of normal tensile micro, intermediate cracks, not accompanied considerable fall of loading on the sample. From $t \sim 1400$ s, and for approx. 20 s, intensive tensile macrocrack generation was observed, with concurrent elimination of strong acoustic impulses that reduce the capacity to resist up to the failure.

This is consistent with the conclusions of Peng (1973) that the load relaxation occurs whenever a crack is open and the magnitude of the load relaxation is proportional to the crack surface. Peng emphasized in particular that the microcracks in the prefailure region are in most cases too small to cause any appreciable load relaxation to be recorded by the load cell. The load dropped sharply at the instant of macrocrack initiation and soon reached certain asymptote values depending on the magnitude of crack surfaces created.

In this context, the peculiarity of EME generation is of specific importance. During the generation of powerful acoustic pulses, EME is absent (under selected threshold of the channels). But in the period directly preceding the burst of strong acoustic pulses (from 1150 to 1300 s), EME are observed, which are not accompanied by a considerable fall of loading or acoustic radiation. The generation of EME in the last phase and its absence in the previous phase could be explained by the higher efficiency of mechano-electric transducers in the case of microcracks which are in most cases too small to cause any appreciable load relaxation to be recorded, and lower efficiency of the transducers in the case of macrocracks. From fig. 6, it follows that the acoustic signals of maximal intensity are recorded directly before and during the fragmentation of the specimen, while EME precedes this stage.

The study of mechanisms of each of the registered types of radiation represents an additional task, but it is possible to ascertain, that EME and AE represent different phases of the destruction process in different efficiency and the combination of both parameters is important for the study of the failure kinetics.

Thus the tests in the condition of self-regulated stress relaxation indicate that EME precedes the phase of macroscopic failure, and is absent during the generation of macrocracks (acoustic pulses), that, on the whole, correspond to *in situ* data of the absence of EME eruption at the onset of the earthquake (Gokhberg *et al.*, 1982, 1995; Yoshino *et al.*, 1992; Slifkin, 1993; Morgounov, 1993a,b, 1996; Morgounov *et al.*, 1994; Hayakawa and Fujinawa, 1994; Varotsos *et al.*, 1996; Vallianatos and Nomikos, 1998; Vallianatos and Tzanis, 1998).

4. Discussion and conclusions

The model of the final stage of the nucleation of the tectonic earthquake has been developed in the paper. The mechanism of tertiary creep under stress relaxation conditions has been assumed as a basis of the interpretation of the short-term precursory events of the impending earthquake. The variety of the precursor images and its duration was explained in the framework of this mechanism. It was emphasized that the onset of tertiary creep appears to correspond to the onset of the instability and inevitably fails unless it is unloaded. At this stage the process acquires to the greatest extent the important property of irreversibility.

Laboratory modeling of the destruction under constant strain and self-relaxed stresses in rocks without supply an external energy (in the condition of the initially accumulated energy) revealed the EME advance with respect to AE activity. Theoretical modeling showed the acceptable correspondence with the field observations of short-lived precursors.

The results outlined above allow us to speculate on some aspects of short-term earthquake prediction. The property of precursors of catastrophic earthquakes to manifest themselves at distances of hundreds and even thousands of kilometers was noted long ago. However, this does not contradict understanding of the process of tectonic earthquake preparation from the standpoint according to which the Earth's crust, comprising a network of faults, is considered a single whole critical dynamic system, in which a change in the stress-strain state at one point results in redistribution of stresses throughout the system.

It seems to be evident in case of geological catastrophes with $M \sim 8$, when planetary-scale processes could involve the upper shells of the Earth. However, it is hardly useful to appeal to the planetary system considering weaker events of $M < 6$ due to the difference of accumulated and released energy. In other words, the magnitude of the earthquake in the first approximation corresponds to the spatial scale of the system of the local faults involved in the active deformation process.

The model of the final stage of nucleation of impending earthquake, discussed above, can be considered the description of the elementary cell of the whole system. During the activation of tectonic activity in the region, some cells of the kind can be developed simultaneously in different critical points of the net of local faults, the dimension of which grows with the magnitude of the earthquake. Moreover, the creep process is developing simultaneously in different stages with different strain rates in different points and affect and interact with one another by the redistribution of tectonic forces that result from the system of the net of faults to the state of self-regulated coalescence (elements of self-organised criticality). Nevertheless, it seems that the study of the elementary process would promote the understanding the mechanism of the interconnection of the parts of a single whole system.

The uncertainty of epicenter location increases with the magnitude of the event, since the zone of $M \sim 7$-8 earthquake preparation amounts to hundreds and thousands of kilometers. The optimum results, therefore, could be expected for the most frequently occurring destructive earthquakes with $M \sim 5$-6. According to Morgounov (2001) the radius of the zone of sensitivity

R^* can be estimated by $R^* = r_f(\varepsilon_f^c/\varepsilon_s^0)^{1/3}$, where: ε_f^c is the failure strain in the focus; ε_s^0 is the strain reference point of start-up of the process near the surface by geophysical parameters. r_f equivalent radius of the focal zone. Using Dambara's formulae for $r_f = 10^{0.5M-2.27}$ (Dambara, 1966) and the strain values ratio: $\varepsilon_s^0/\varepsilon_f^c \sim 10^{-6}$ we obtain for the maximal ($\varepsilon_s^0 \sim 10^{-9}$) and effective (for instance, $\varepsilon_s^0 \sim 10^{-6}$) radii of the preparation zone respectively: $R^*_{max} \sim 10^{0.5M-0.27}$ km and $R^*_{eff} \sim 10^{0.5M-1.27}$ km.

Laboratory tests show that the more homogeneous the medium is, the greater the difficulty in identifying the diagnostic features of its fracture, and *vice versa:* the more inhomogeneous is the specimen, the earlier the stage at which fracture precursors appear (Shamina, 1981). Thus the heterogeneity of the Earth crust could be in a way a favourable factor for timely distinction of the precursors of the shock *in situ* condition.

Kranz (1979) noted that stresses within the Earth generally do not vary rapidly. Thus the quasi-static state of stress in the Earth is more closely modelled by creep tests than by a constant strain-rate test in which stress is continuously incremented. The tests of rocks under the load relaxation condition described above show that two important phases could be distinct immediately before failure.

The impulsive EM emission is predominated during the first phase because of opening micro-intermediate size of cracks but the strong acoustic pulses are predominated during the second (failure) phase, accompanied by the relaxation of load up to the unloading. Moreover, powerful AE during the second phase were not accompanied by concurrent EM radiation (at least as powerful as in the previous phase). The absence of strong EME can be explained by lower efficiency of the mechano-electric transducer mechanism in case of macrocracks. The laboratory results basically correspond to *in situ* data. EME is usually observed immediately before the earthquake and there is no burst of radiation on the onset of the earthquake (co-seismic anomaly is of the order of magnitude of pre-seismic one).

The experience of applied studies of electromagnetic precursors in different countries showed that a complex of electromagnetic events appears tens of hours before an earthquake. The long term observations at a certain region enables us to eliminate the local noise and to find the repeating precursor signature that enables us to develop the algorithms for their identification and to realize the regular scientific short-term earthquake prediction of local earthquakes of $M < 4.5$ (Zdorov and Morgounov, 1997).

The remarkable feature of the terminal stage of earthquake nucleation is that once a rock enteres this stage it inevitably fails until it is unloaded. At this stage, the process acquires the self-regulating character and to a great extent the property of irreversibility, one of the important components of prediction reliability. *In situ*, the load conditions change during a considerable inelastic deformation that results in the multifaced precursor signature including oscillatory precursor and false alarm.

Acknowledgements

I sincerely thank the Organising Committee and Acc. Prof. Filippos Vallianatos for the kind invitation to attend the «2nd International Workshop on Magnetic, Electromagnetic and Electromagnetic Methods in Seismology and Volcanology», Chania, Greece, September 22-24, 1999. I am gratefully indebted to Dr. Andreas Tzanis whose reasonable critical comments and corrections of English helped me to improve the clarity of the paper.

REFERENCES

BENIOFF, H. (1951): Earthquakes and rock creep, *Bull. Geol. Soc. Am.*, **41** (1), 31- 62.

EVANCE, J.R. and D.C. BOOTH (1996): Assessment schemes for earthquake prediction, Astronomical Society/Joint Association for Geophysics, *Discussion Meeting*, 7 and 8 November 1996 at the Geological Society, Burlington House, Piccadilly, London [http://www.ras.org.uk/ras].

BRADY, B.T. (1976): Theory of earthquakes (4), *Pageoph*, **114** (6), 1031-1041.

CORWIN, R.F. and H.F. MORRISON (1977): Self-potential variations preceding earthquakes in Central California, *Geophys. Res. Lett.*, **4**, 171-174.

DAMBARA, T. (1966): Vertical movements of the Earth's crust in relation to the Matsushiro earthquake, *J. Geod. Soc. Japan*, **12**, 18-45.

GELLER, R.J. (1997): Earthquakes: thinking about the unpredictable, *Eos*, February 11, p. 67.

GLUSHKO, V.T. and V.V. VINOGRADOV (1982): *Rock Fracture and Prediction of Rocks Bursts* (Moscow, Nedra), p. 191.

GOKHBERG, M.B., V.A. MORGOUNOV, T. YOSHINO and I. TOMIZAWA (1982): Experimental measurements of electromagnetic emissions possibly related to earthquakes in Japan, *J. Geophys. Res.*, **82**, 7824-7888.

GOKHBERG, M.B., V.A. MORGOUNOV and O.A. POKHOTELOV (1995): *Earthquake Prediction. Seismo-electromagnetic Phenomena* (Gordon and Breach Publisher), p. 191.

GRIGGS, D. (1940): Experimental flow of rocks under conditions favoring reclystallization, *Bull. Geol. Soc. Am.*, **51** (2), 1001-1022.

HAYAKAWA, M. and Y. FUJINAWA (Editors) (1994): *Electromagnetic Phenomena Related to Earthquake Prediction* (Terra Sci. Pub. Co., Tokyo), p. 677.

KENNEDY, A.J. (1962): *Processes of Creep and Fatigue in Metals* (Oliver and Boyd, Edinburgh and London), p. 312.

KRANZ, R.L. (1979): Crack growth and development during creep of Barre granite, *Int. J. Rock Mech. Min. Sci. Geomech.*, Abstr., **16**, 23-35.

KRANZ, R.L. and C.H. SCHOLZ (1977): Critical dilatant volume of rocks at the onset of tertiary creep, *J. Geophys. Res.*, **82** (30), 4893-4898.

LOCKNER, D. and J. BYERLEE (1980): Development of fracture planes during creep in granite, in *Second Conference on Acoustic Emission/Microseismic Activity in Geologic Structures and Materials - The Pennsylvania State University, November, 13-15, 1978* (Trans. Tech. Pub., Clausthal, Germany), 11- 25.

MORGOUNOV, V.A. (1993a): Creep in geodynamic processes, *Trans. Russian Academy of Sciences/Earth Science Sections*, **317** (6), 1343-1352 (English translation).

MORGOUNOV, V.A. (1993b): Real-time earthquake forecasting, *Trans. Russian Academy of Sciences/Earth Science Sections*, **319** (1), 138-142 (English translation).

MORGOUNOV, V.A. (1996): Tertiary creep in focal zone and immediate time earthquake prediction, *J. Earthquake Prediction Research*, Beijing, Moscow, **5** (3), 155-181 (English translation).

MORGOUNOV, V.A. (2001): Rock creep at the final stage of earthquake nucleation, *Izv., Phys. Solid Earth*, **37** (4), 279-286 (English translation).

MORGOUNOV, V.A., M.N. LUBASHEVSKY and Z.E. FABRIZIUS (1991): Geoacoustic precursor of Stitak earthquake, *Volcanol. Seismol.*, **4**, 104-106.

MORGOUNOV, V.A., T. ONDOH and S. NAGAI (1994): Anomalous variations of VLF signals associated with strong earthquakes ($M > 7.0$), in *Electromagnetic Phenomena Related to Earthquake Prediction*, edited by M. HAYAKAWA and Y. FUJINAWA, Tokyo, 409-428.

MORGOUNOV, V.A., YU. V. KULINICH and A.A. SIROTIN (2001): Electromagnetic and acoustic emission at the final stage of rock fracture, *Earthquake Prediction J.*, **9** (3), 285-293.

OGAWA, T., K. OIKE and T. MIURA (1985): Electromagnetic radiation from rocks, *J. Geophys. Res.*, **90** (D 4), 6245-6249.

OHNAKA, M. (1983): Acoustic emission during creep of brittle rock, *Int. J. Rock Mech. Min. Sci. Geomech. Abstr.*, **20** (3), 121-134.

PENG, S.S. (1973): Time-dependent aspects of rock behaviour as measured by a servocontrolled hydraulic testing machine, *Int. J. Rock Mech. Min. Sci. Geomech., Abstr.*, **10**, 235-246.

RICE, J.R. and J.W. RUDNICKI (1979): Earthquake precursory effects due to pore fluid stabilisation of weakening fault zone, *J. Geophys. Res.*, **84** (B5), 2177-2193.

SCHOLZ, C.H. (1968): Mechanism of creep in brittle rock, *J. Geophys. Res.*, **73** (10), 3295-3302.

SHAMINA, O.G. (1981): *Model Studies of the Earthquake Source Physics* (Moscow, Nauka), p. 191.

SLIFKIN, L. (1993): Sesmic electric signals from displacements of charged dislocations, *Tectonophysics*, **224**, 149-152.

TAPPONIER, P. and W.F. BRACE (1976): Development of stress-induced microcracks in Westerly granite, *Int. J. Rock. Mech. Min. Sci.*, **13**, 103-112.

VALLIANATOS, F. and A. TZANIS. (1998): Electric current generation associated with the deformation rate of a solid: preseismic and coseismic signals, *Phys. Chem. Earth*, **23** (9-10), 933-938.

VALLIANATOS, F. and K. NOMIKOS (1998): Seismogenic radioemissions as earthquake precursors in Greece, *Phys. Chem. Earth*, **23** (9-10), 953-957.

VAROTSOS, F., K. EFTAXIAS, F. VALLIANATOS and M. LAZARIDOU (1996): Basic principles for evaluating an earthquake prediction method, *Geophys. Res. Lett.*, **23** (11), 1295-1298.

WARWICK, J.W., C. STOKER and T.R. MAYER (1982): Radio emission associated with rock fracture: possible application to great Chilean earthquake of May 22, *J. Geophys. Res.*, **4**, 2851-2859.

WU, F.T. and L. THOMSEN (1975): Microfracturing and deformation of Westerly granite under creep condition, *Int. J. Rock Mech. Mining Sci. Geomech. Abst.*, **12**, 167-173.

YOSHINO, T., I. TOMIZAWA and T. SIGIMOTO (1992): Results of statistical analysis of LF seismogenic emissions as precursors to the earthquake and volcanic eruptions, *Res. Lett. Atmos. Elect.*, **12**, 203-210.

ZDOROV, A.G. and V.A. MORGOUNOV (1997): An experience of short-term earthquake prediction using electromagnetic radiation technique: a Mineralnye Vody, Caucasus, case study, *Trans. Russ. Acad. Sci./ Earth Sci. Sect.*, **357** (8), 1257-1260 (English translation).