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On the Long-Range Influence of Earthquake Rupture Zones

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Abstract—This paper proposes a new approach to estimating the long-range influence of crustal volumes that involve anomalous stress behavior. It is shown that the consideration of this problem within elasticity theory underestimates the long-range action, determining a rapid decrease in the influence of the volume involving anomalous stresses on the state of stress in the earth crust, as one goes away from the source of the anomaly. In these approaches the influence of the anomaly at great distances (100 radii of the anomalous volume) is due to a reduction of strain by a factor of 10^6 compared with the deformation in the anomalous volume. It is proposed in estimating the long-range influence of anomalous volumes to take account of the fact that there are extensive regions in a supercritical state at different crustal horizons. Such regions are fault zones in the upper crust and layers of high fluid pressure in the middle crust (waveguides). The presence of regions of supercritical deformation where the earth behaves inelastically (pseudoplastic and quasiplastic behavior) produces a different regime of long-range influence due to incipient anomalies of stress, different from the purely elastic response. We showed that in the 2-D case (a plane horizontal layer) and the classical Coulomb–Mohr criterion, the reduction in the stress disturbance due to an anomalous inclusion can be given by the law $1/r$ (r is the lateral distance relative the center of the inclusion when normalized by the size of the inclusion). The expressions derived in this study show that the level of disturbed strain at great distances is reduced by a factor of only 100 when under horizontal compression or tension. This is lower by a factor of 10^4 compared with what would be expected in an elastic medium. Consequently, the change in the level of stress due to an anomaly in a medium that is experiencing irreversible deformation occurs at a substantially lower rate in relation to increasing distance than in a purely elastic medium, where the decay of stress behaves like $1/r^2$.

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INTRODUCTION

In the theory of earthquake sources the problem occurs of how it is possible to find the long-range influence, if any, of an imminent earthquake on physical fields in the crust at great epicentral distances. This problem has become urgent in connection with the study of earthquake precursors, which total over 300 at the present time (Sobolev, 1993; Sidorin et al., 1992). It is generally thought that an earthquake source (ES) is a region where the stresses are at an anomalous level relative to the background stresses around it. Depending on the type of an inclusion, hard or soft, one distinguishes different concepts of ES precursory process (Pevnev, 2003). This problem also exists in application to magma-conducting conduits and deep magma chambers beneath volcanoes.

According to Reid's elastic rebound theory (Reid, 1910), an earthquake is preceded by the appearance of a locked patch on the responsible fault; this patch causes a gradual buildup of near-fault stresses up to a critical value. Starting from this concept, a precursory region must be viewed as a *hard inclusion* where the energy of elastic strain is accumulated, hence the level of stress increases (Benioff, 1951; Bullen, 1953; Gzovsky, 1957, 1975; Gamburtsev, 1960; Ulomov and Mavashev, 1967; Riznichenko, 1968). The time this elastic energy takes to accumulate is related to the magnitude of the future earthquake (Sadovsky and Pisarenko, 1985).

The ES theory according to Richter (1963) differs from Reid's theory in that the strain and stress are assumed to be relatively uniformly distributed in space prior to an earthquake. As a matter of fact, this explanation of earthquake occurrence involves the hypothesis of a local diminution of strength and mechanical properties on a fault patch, which determines the possibility of a reduced stress level in the source region of the future earthquake, with the precursory region being a *soft inclusion*.

The additional stresses that occur in the region of the future earthquake are changing with increasing distance from it and are gradually approaching the background values (they decrease for the case of a hard inclusion and increase for a soft). In this way there is a relationship between the intensity of an earthquake precursor and the distance from the earthquake source where that precursor can be detected (Sidorin, 1979). Several researchers have pointed out that a long-range action such as this does not occur everywhere. The crust contains special volumes (faults and strain-sensitive zones, Kissin, 2009) that can give rise to large changes in physical parameters.

The fact that the Richter model is valid for the development of an ES is corroborated by quite recent data on patterns in stress distribution in the precursory regions of great earthquakes (Rebetsky and Marinin, 2008; Rebetsky, 2009). These studies were based on the method of cataclastic analysis of discontinuous displacements (MCA) which was developed at the Laboratory of Tectonophysics, Institute of Physics of the Earth RAS (Rebetsky, 1999, 2001, 2003, 2005, and 2009). This method is an extension of known tectonophysical methods for determination of tectonic stresses due to O.I. Gushchenko (1975, 1979), S.L. Yunga (1979, 1990), and J. Anjelier (1975, 1984), but, in contrast to these earlier methods, ours can estimate the stresses themselves, in addition to the orientations of principal stress axes. To do this, as proposed by Anjelier (1989), the MCA incorporates results from experiments on rock failure (Byerlee, 1978; Brace, 1978; Stavrogin and Protosenya, 1992), as well as additional seismological data on dynamic source parameters (stress drops) for the larger earthquakes that have occurred in an area of study.

The results from tectonophysical surveys in the areas of several great earthquakes (Sumatra–Andaman $M = 9.1$, 2004; Simushir $M = 8.1$, 2006; and Chilean $M = 8.9$, 2009) have firmly established the fact that the source region of a future large earthquake is formed in a crustal volume under a lower level of effective confining pressure. The lower level of effective pressure also determines a lower level of shear stresses required for activating the rupture (a displacement at the earthquake source), as well as a lower level of the elastic energy that is expended to overcome the frictional resistance. The size of the volume actually determines the magnitude of the future earthquake. The rupture is initiated either where the gradient of effective pressure is at the maximum at the boundary of the future earthquake source or in a volume where the effective pressure inside the source volume is locally higher (Rebetsky and Marinin, 2006a, b; Rebetsky, 2007a, b).

To sum up, contemporary knowledge concerning the state of stress in regions of earthquake precursory processes characterizes a future source as a zone where the stress level experiences a sharp change in the form of a soft inclusion.

A few words are in order to clarify the terminology. An earthquake source is understood as a fault plane along which a shear displacement occurs at the time seismic waves are initiated, which is the start of an earthquake (the source is a two-dimensional region). The ES precursory region is here understood as the surrounding 3-D volume where the stress change occurs with the maximum intensity. The precursory region is the volume that radiates seismic energy. We note that Bullen (1953) understood the ES as the region that released elastic energy during the movement at the source.

The term stress, when used without refining definitions, means both the average pressure that is responsible for the deformation change in rock volume and the deviatoric (differential) stresses, whose main changes occur during the stress drop at the earthquake source due to shear displacements at its sides. Both of these components of the stress tensor are controlled by the weight of the overlying rocks (vertical lithostatic pressure) and an additional tectonic component that is caused by movement in adjacent rock volumes.

THE REGION OF INFLUENCE DUE TO AN EARTHQUAKE SOURCE

Working on the assumption of an ES precursory region as an anomalous inclusion, Dobrovol'sky et al. (1979), Dobrovol'sky (1991) have developed mathematical tools that can be used to find the level of change in disturbed stress as the distance from the anomalous inclusion increases. They have solved the mechanical problem of the influence of an anomalous inclusion on the state of stress in an elastic half-space. It was thought that an ellipsoidal region is formed in a homogeneous field of initial horizontal compressive stresses (*anomalous inclusion*) where the elastic properties change. If the elastic moduli in the region increase relative to their initial values (the moduli in the space around the inclusion), then we have a hard inclusion in a half-space (a consolidation model of the earthquake precursory process); if they decrease, the inclusion is soft. It was desired to find the disturbed state of strain and stress, which was found as the difference between the state in the presence of the inclusion and the initial homogeneous state. It was assumed when constructing the solution that the state of stress given by the force of gravity is isotropic, i.e., it produces the same level of compressive stresses in all directions equal to lithostatic pressure p_{lt}. These stresses were disregarded in the subsequent solution.

The deformations outside the inclusion were found using the theory of small disturbances, which reduces the solution to a volume integral of Green's function for certain *fictitious forces* that act at points within the anomalous inclusion. The magnitudes and directions of these forces are determined by those increments to the impulse of force, which arise at the initial state of stress due to changes in the elastic moduli in the inclusion. A solution was derived for small changes in the elastic parameters that should be treated as an approximate one for the far zone of influence of the inclusion. Several situations were considered for the initial state of stress: pure shear in the horizontal plane and a uniaxial compression in the horizontal plane. Below, we quote a formula from (Dobrovol'sky, 1991, p. 109) for estimating the change in the modulus of the maximum principal longitudinal strain (ε_d) at the surface of the half-space outside the epicentral area for the case of a spherical inclusion of radius R as derived for the initial state of pure shear ($\sigma_1 = \tau$, $\sigma_2 = 0$, $\sigma_3 = -\tau$) for the geodynamic regime of horizontal compression $(\sigma_{zz} = \sigma_2)$:

$$\varepsilon_d = \varepsilon_a \left(\frac{R}{r}\right)^3, \text{ when } \varepsilon_a = \frac{31\pi^2}{12}\alpha\gamma_0, \quad \gamma_0 = \tau/\mu. \tag{1}$$

Here, γ_0 is the maximum elastic shear strain at the initial state of stress, μ and $\alpha \approx 0.1$ are the modulus of elastic shear and a constant that determines the change in the shear modulus in the anomalous region $(\alpha\mu)$, and the τ are the maximum shear stresses at the initial state of stress. It was assumed for this calculation that $\mu = 2\times10^{10}$ Pa, $\tau = 10^8$ Pa, from which we obtained the elastic strain released at the earthquake source, $\varepsilon_a \approx 1.27\times10^{-2}$.

We wish to draw attention to the fact that the magnitude of strain decreases according to the law $(R/r)^3$ as the distance from the anomalous inclusion increases. This is a typical pattern in the framework of elasticity theory for a 3-D stress distribution. When the stress field is two-dimensional, the law becomes quadratic: $(R/r)^2$.

It was proposed to estimate the deformation influence of an anomalous inclusion (the region of an incipient earthquake source) on the surrounding space by introducing the lowest level of strain $\varepsilon_d = 10^{-8}$, that we consider, corresponding to the maximum level of tidal strain. I.P. Dobrovol'sky dealt with the ES precursory process using a consolidation model. The concept implies that the variation of physical processes that are occurring in a future earthquake source ultimately precludes the dissipation of mechanical energy pumped into the earth via plastic and discontinuous deformations (the appearance of an asperity). The consequence is that the energy of elastic strain and corresponding stresses begin to increase in that region (the Reid model for precursory processes).

According to (1), the elastic strain that takes place at the boundary of the inclusion $\varepsilon_a \approx 0.013$, reaches $\varepsilon_d = 10^{-8}$, at the boundary of the area of influence, that is, is reduced by a factor of nearly 10^6.

From (1) we can derive a relation connecting the size of the precursory area (R) and the radius of its long-range influence (r_d):

$$r_d/R \approx 110, \tag{2}$$

which causes precursors of different types to occur (Sidorin, 1979).

We note that the approach of Dobrovol'sky to the estimation of the long-range influence exerted by precursory processes can also be applied to the problem of detecting changes in the mode of supply in the magma-conducting conduit (chamber) beneath a volcano. The last section of this paper will interpret the results as applicable to this problem as well.

A DISCUSSION OF ESTIMATES FOR THE LONG-RANGE INFLUENCE OF AN EARTHQUAKE SOURCE

The idea of Dobrovol'sky on finding the boundary of the area of inclusion-induced influence based on the strain value on the order of 10^{-8} calls for several remarks. In the first place, it is commonly considered in mechanics that the accuracy of stress calculation must be consistent with the accuracy of the data on elastic parameters and strength. In application to geosciences, the possible variations in the elastic moduli are 100% and still greater. As an example, in seismology (rapid strain change) it is supposed that the elastic shear modulus for rocks in the consolidated crust has values on the order of $3–5 \times 10^{10}$ Pa. At the same time, the effective elastic modulus for bedrock as used in mining mechanics, with due incorporation of porosity and cracking, can reach values of the order of $0.1–1 \times 10^{10}$ Pa (Vlasov and Merzlyakov, 2009). Secondly, the accuracy of stress calculation is determined by the degree of approximation inherent in the mathematical tools we are using in the mechanics of solids. Thirdly, in application to ES, the accuracy of calculation is determined by how accurate the geometry of the inclusion is that gives rise to the stress anomaly.

All the above tells us that when a strain on the order of 10^{-8} is used to estimate the long-range distance, small variations in elastic properties, under which strain is incorporated or refined in the used mathematical tools (incorporation of quadratic forms of strain), can decrease or increase the distance by a factor of many times. The long-range distance will be small if the level of disturbed strain is increased to reach $\varepsilon_d = 0.01\varepsilon_a$ (0.01 is the highest accuracy that is attainable in determinations of the elastic shear modulus in rock massifs). In that case the desired level of disturbed strain will be on the order of 0.0001, so that, in accordance with (1), the size of the area of influence will be diminished by a factor of nearly 20. The ratio will change accordingly: $(r/R) \approx 5$. We note that this relation is consistent with the reasonable dimensions of areas where anomalous inclusions affect the results as adopted in structural mechanics.

The level of limiting strain chosen in (Dobrovol'sky, 1991) to calculate the size of the long-range area was chosen on the basis of the accuracy attainable in the identification of useful signals in noise, which is here meant to be lunar tidal strains. However, a strain level on the order of 10^{-8} cannot give rise to phenom-

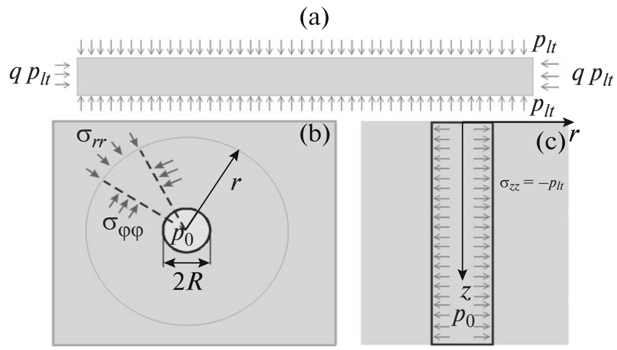

Fig. 1. A scheme showing the problem of stresses in a plane layer within a massif with a vertical inclusion in the layer in the shape of ac cylinder with a constant pressure p_0 specified at the cylinder walls. (a) the position of the layer within the massif at the initial state of stress; (b, c) view of the inclusion in the shape of a vertical cylinder in map view (b) and in a cross section (c). R is the radius of the inclusion and r is the distance between the point of interest and the origin of coordinates.

ena that would be classified as precursors (e.g., changes in water level in wells of a few meters or a few tens of meters). Otherwise, lunar tidal strains would certainly create such anomalies.

However, the chief argument that emphasizes the contradictions in the concept of calculating the area of influence according to I.P. Dobrovol'sky consists in the fact that the geological medium is in a supercritical state before any earthquake source begins to originate. A supercritical state is meant here as the stresses that act in a rock massif that reaches some limiting relation that determines the generation of irreversible deformation in the rocks. In application to the rocks in the upper and middle crust, we are concerned, not with the yield limit and the appearance of genuine plastic deformation (dislocations in grains and crystals) (Nikolaevsky, 1996), but with crack deformation, which occurs when the cataclastic yield limit has been exceeded (Rebetsky, 2007a).

Rebetsky (2008a, b) showed that for an initial gravitational state of stress (this is invariably present before external deforming forces have been applied) the cracking yield limit can be reached at depths as shallow as 1–2 km in consolidated rock when its cracking is high (a lower effective cohesion) and there is a strength-reducing fluid pressure after K. Tertsagi (1961). This higher cracking is typical of large regional and interregional fault zones where all large earthquakes occur. Outside of fault zones the medium that has a higher effective cohesion will attain a supercritical state at great depths (3 to 10 km), which depends on upper crustal tectonics.

The presence of background seismicity that covers a large area of a seismic region with small and moderate earthquakes (M = 0.5–2) shows that the crust in seismic regions is beyond the cataclastic yield limit everywhere. When such background seismicity is nearly absent and all seismic events occur on fault zones and in their immediate vicinities, it is only the rocks in these fault zones that are beyond the yield limit.

Cataclastic flow in rock massifs is controlled by activation of a set of brittle cracks and looks like plastic flow due to the averaging scale of crack deformation (pseudoplasticity). In application to the description of such a flow in regional and interregional crustal fault zones, the maximum linear dimensions of the cracks that give rise to such cataclastic flow are a few tens to a few hundreds of meters (the width of a fault zone can reach a few tens of kilometers). In this case, cracks and faults a few kilometers long or longer should be regarded as an act of brittle fracture that is developing in a medium beyond the pseudoplastic (cataclastic) yield strenth.

AN APPROACH TO THE ESTIMATION OF THE AREA OF INFLUENCE DUE TO AN INCIPIENT EARTHQUAKE SOURCE

Understandably enough, there is a wide difference in the stress distribution around an inclusion that gives rise to a stress anomaly between the cases in which the medium is in a purely elastic state and when it is elastoplastic. In the latter case its ability to resist the extra stress around the anomaly would be limited. We shall consider the problem of an anomalous inclusion in a massif for two different cases of its response to strain: (1) purely elastic strain and (2) elastoplastic strain. In the latter case the massif is assumed to have reached a supercritical state before the formation of an anomalous inclusion, when it was only under the initial stress.

The simplest way to discern the difference between the two states is by considering the example of a 2-D problem for a band with a cylindrical inclusion with pressure p_o specified on the walls of the inclusion (Figs. 1b, c). We must cut a layer out of the massif on whose horizontal faces we specify a vertical pressure $p_{lt} = \rho g z$ (ρ is the density of the material, g is the acceleration due to gravity, and z is the depth) such as would be caused by the weight of the overlying rocks and at infinity we specify the lateral resistance stresses qp_{lt} (see Fig. 1a) where q is the coefficient of lateral resistance. We want to find the function describing the variation of stress beyond the cylinder walls. We considered different relationships between the lithostatic pressure p_{lt} and the pressure p_o, that acts on the walls of the cylindrical inclusion.

It is important to mention that problems of this kind have been successfully handled when studying the stability of walls in a well (Schmitt et al., 2012) for the development of hydrocarbon deposits. There are many theoretical estimates of the stress in the layer beyond a well, both in the elastic and in the elastoplastic formulation (Yu and Rowe, 1999; Xu, 2007). The present study adapts known solutions to the problem of distant action due to dangerous geological processes such as earthquakes and volcanic activity.

The initial state of stress. We begin by considering the state of the layer without the cylindrical inclusion (see Fig. 1a). We assume the initial state to be the sum of the gravity pressure in a layer at a depth, the body forces, and the extra pressure due to horizontal compressive stresses that give rise to lateral normal stresses that have equal magnitudes (the lateral resistance stresses):

$$\sigma_{zz}^0 = -p_{lt},\ \sigma_{rr}^0 = \sigma_{\varphi\varphi}^0 = -qp_{lt}, \tag{3}$$

where σ_{zz}^0, σ_{rr}^0, and $\sigma_{\varphi\varphi}^0$ are the initial values of normal stresses in cylindrical coordinates (r, φ, z). Here and below, the superscript 0 denotes an initial laterally homogeneous state of stress that is only controlled by gravity forces or, in cases to be specified below, also by the extra stresses of horizontal compression.

We shall neglect the changes in stress due to body forces within the layer itself, that is, we assume the layer thickness to be small compared with the depth of the layer.

Depending on the level of lateral resistance, the geodynamic type of stress also changes. If (1) $\sigma_{\varphi\varphi} = \sigma_{rr} > \sigma_{zz}$ $(0 < q < 1)$, then we have the mode of horizontal tension with the stress tensor for uniaxial compression. In case (2) $\sigma_{zz} > \sigma_{\varphi\varphi} = \sigma_{rr}$ $(q > 1)$ we have the mode of horizontal compression with the stress tensor for uniaxial tension. For $q = 1$ the medium is free of deviatoric stresses and only normal stresses equal to the lithostatic pressure act on differently oriented inclined areas.

When the medium undergoes supercritical deformation, the coefficient of lateral resistance depends on the yield parameters at the phase of initial state of stress when the limit of pseudoplastic (cataclastic) flow has been exceeded. We shall use the Coulomb–Mohr criterion in what follows as the limiting relation for the elastoplastic transition:

$$\tau = \tau_c - k_c(\sigma_\tau + p_{fl}) \text{ when } \tau = (\sigma_1 - \sigma_3)/2 \\ \text{and } \sigma_\tau = -(\sigma_1 + \sigma_3)/2. \tag{4}$$

Here, τ_c and k_c are the the limit of internal cohesion strength and the limit of internal friction in rocks, τ is the maximum shear stress, σ_τ is the normal stress on the plane of the maximum shear stress, p_{fl} is the fluid pressure in the fissure and pore space that diminishes the friction in fissures, while $\sigma_1$and σ_3 are the algebraically greatest and the lowest principal stress, respectively (tension is positive).

We note that the limiting relation was written somewhat differently in (Rebetsky and Lermontova, 2016), viz., instead of σ_τ we used the mean of the three principal stresses, or isotropic pressure. This modification to the Coulomb–Mohr criterion was an attempt to incorporate the third principal stress.

The subsequent calculation will use the value p_{lt} = 25 MPa, which is relevant to the depth of the top of the layer at approximately 1 km, as the stress that acts on the horizontal boundaries of the layer. The strength parameter values will be $\tau_c = 2.5$ MPa, $k_c = 0.5$, and $\lambda = p_{fl}/p_{lt} \approx 0.4$ (the fluid pressure is approximately the weight of the water column).

Using (4) for the geodynamic regime of horizontal tension $(\sigma_{\varphi\varphi} = \sigma_{rr} > \sigma_{zz})$, we find:

$$q = -\frac{2\tau_c/p_{lt} - 1 + k_c(1 - 2\lambda)}{1 + k_c} \text{ with } \frac{\nu}{1-\nu} < q < 1, \tag{5}$$

where ν is Poisson's ratio. The left-end bound for q is given by the value of lateral resistance during the elastic phase after (Dinnik, 1926). Calculation based on the above values of strength and lithostatic pressure yields $q \approx 0.467$, while the horizontal compressive stresses are $\sigma_{\varphi\varphi}^0 = \sigma_{rr}^0 = -0.467p_{lt} = -11.7$ MPa.

We wish to note that the requirement for lateral resistance q imposes certain restrictions on the loading and strength parameters. In particular, when τ_c, p_{lt} and λ, are specified, the frictional constant (k_c) cannot be arbitrary. In other words, all of these parameters are connected through (4).

We assume that the stresses proper to the regime of horizontal compression are supercritical as well. Using (4) for this type of geodynamic regime $(\sigma_{zz} > \sigma_{\varphi\varphi} = \sigma_{rr})$, we find

$$q = \frac{2\tau_c/p_{lt} + 1 + k_c(1 - 2\lambda)}{1 - k_c} \text{ when } q > 1. \tag{6}$$

Further, we find $q \approx 2.6$ from (6) for the above strength parameters and the value of lithostatic pressure. The horizontal compressive stresses $\sigma_{\varphi\varphi}^0 = \sigma_{rr}^0 = -2.6p_{lt} = -65$ MPa thus correspond to the elastic phase for this depth.

We note one important issue. The restrictions on lateral resistance in (5) and (6) lead one to a restriction on the choice of parameters that determine q during the phase of elastoplastic deformation.

We now discuss the change in the state of stress, if a cylindrical inclusion appears in the layer with lateral pressure $p_0 > 0$ at its walls.

The solution of the elastic problem. We begin by considering the case in which the medium responds purely elastically to the pressure exerted by a cylindrical inclusion (see Fig. 1). Since the problem is now axisymmetric $(\sigma_{r\varphi} = \sigma_{z\varphi} = \sigma_{rz} = 0)$, the resolvent equations include an equilibrium equation (Rabotnov, 1979)

$$\frac{\partial \sigma_{rr}}{\partial r} + \frac{\sigma_{rr} - \sigma_{\varphi\varphi}}{r} = 0, \tag{7}$$

Cauchy's relations, and Hooke's law. This is a square set of equations, that is, the number of unknowns and the number of equations are equal.

We shall consider a thin plane layer within a half-space where the change of initial gravitational stress state is small compared with the stress level in the middle plane of the layer. We assume lithostatic pressure p_{lt} to be acting. In this problem we will not consider the interaction of the layer with the overlying and underlying layers, assuming the strain differences to be small.

When the pressure p_0 is constant at the contact of the anomalous region with the outer massif, the solution of the elastic problem is (see, e.g., (Rabotnov, 1979))

$$\begin{aligned} \sigma_{rr} &= -qp_{lt} + (qp_{lt} - p_0)(R/r)^2, \\ \sigma_{\varphi\varphi} &= -qp_{lt} - (qp_{lt} - p_0)(R/r)^2. \end{aligned} \tag{8}$$

Here, q is the coefficient of lateral resistance for an initial state without an inclusion. The construction of the elastic solution requires that the initial gravitational state be elastic and the value of lateral resistance should be given by a relationship after A.N. Dinnik (5) or (6).

We wish to note that the solutions (8) are for the case in which the lateral stresses are independent of the polar angle θ, which is measured from any radial direction. It is important to note that the decay of disturbing stresses in the resulting expressions (the second term in radial and tangential stresses) is inversely proportional to the square of the radius, which is consistent with I.P. Dobrovol'sky's model for the case of a two-dimensional state of stress. When the solution proper for non-isotropic lateral stresses $\left(\sigma^0_{xx} \neq \sigma^0_{yy}\right)$, is used, extra terms appear in (8) (see Kirsch, 1898) that are proportional to the difference between the lateral stresses and determine the decay of the disturbance according to the law $(R/r)^2$ and $(R/r)^4$.

Further, when we consider various cases of stress around a cylindrical inclusion with a constant pressure on its walls, we shall also vary the relationship between the pressure at the walls and the level of lateral stresses, in addition to varying the type of initial gravitational state of stress (horizontal compression or tension). Viewed from the standpoint of the theory that describes the precursory process of an earthquake, or of changes in the regime of the magma-conducting volcanic conduit, when the pressure that arises in the inclusion is below the level of lateral pressure due to the initial gravitational state of stress $(\sigma_{\varphi\varphi} + \sigma_{rr})/2$, we can consider the cylindrical inclusion to be a *soft inclusion*. Accordingly, if the pressure that is created in the inclusion is above the initial lateral pressure, then the cylindrical inclusion can be viewed as a *hard inclusion*.

The formulation of the elastoplastic problem. We consider the case in which the layer, in the initial state, is beyond the yield limit. One of the principal stresses is known, similarly to the case of the elastic phase. This is the vertical stress $\sigma_{zz} = -p_{lt}$. The other two unknown stresses $\sigma_{\varphi\varphi}$, σ_{rr} can be found from two equations: the equation of equilibrium (7) and the Coulomb–Mohr criterion of failure (4).

We wish to note that the approach we are using to find an analytical solution to the plasticity problem is not new. There is a rather simple solution for the case of an isotropic state of stress in the lateral direction to be found in many textbooks (Rabotnov, 1979), which is relevant to the transition to a supercritical state following the Tresca model (the maximum shear stress is the limit). For this criterion for transition to plasticity, there are also solutions for the case in which the initial lateral stresses are unequal. This is L.A. Galin's problem, and the resulting solutions were used in structural mechanics (Savin, 1968). There are quite recent publications with solutions for a similar problem in the Coulomb–Mohr supercritical model, which were used to predict the behavior of wells and tunnels (Yu and Rowe, 1999; Xu, 2007).

The special feature in the use of these solutions for the problem under consideration here consists in incorporation of different geodynamic types of stress state and different relationships between the pressure in the inclusion and the middle lateral stress in the initial state.

Our solution will make use of the fact that normal stresses that act in rock massifs are generally little different from lithostatic pressure $(\sigma_{zz} = -p_{lt})$, while the change in the geodynamic type of the state of stress is due to a change in lateral stresses.

The level of pressure existing in an inclusion can affect, not only the relationship between the radial and the tangential stress, but also that between these stresses and the vertical stress σ_{zz}. There are six variants of possible sets of relationships between these stresses, as follows: Ia) $\sigma_{\varphi\varphi} > \sigma_{rr} > \sigma_{zz}$, Ib) $\sigma_{rr} > \sigma_{\varphi\varphi} > \sigma_{zz}$, IIa) $\sigma_{zz} > \sigma_{rr} > \sigma_{\varphi\varphi}$, IIb) $\sigma_{zz} > \sigma_{\varphi\varphi} > \sigma_{rr}$, IIIa) $\sigma_{rr} > \sigma_{zz} > \sigma_{\varphi\varphi}$, IIIb) $\sigma_{\varphi\varphi} > \sigma_{zz} > \sigma_{rr}$. According to the above inequalities, the first two are proper to horizontal tension, the next two to horizontal compression, and the last two to horizontal shear. We shall examine which of these variants can occur in the two cases of initial state of stress as specified by the level of lateral resistance (5) and (6).

State Ia: horizontal tension, a hard inclusion. We assume the pressure level in the inclusion to be such as

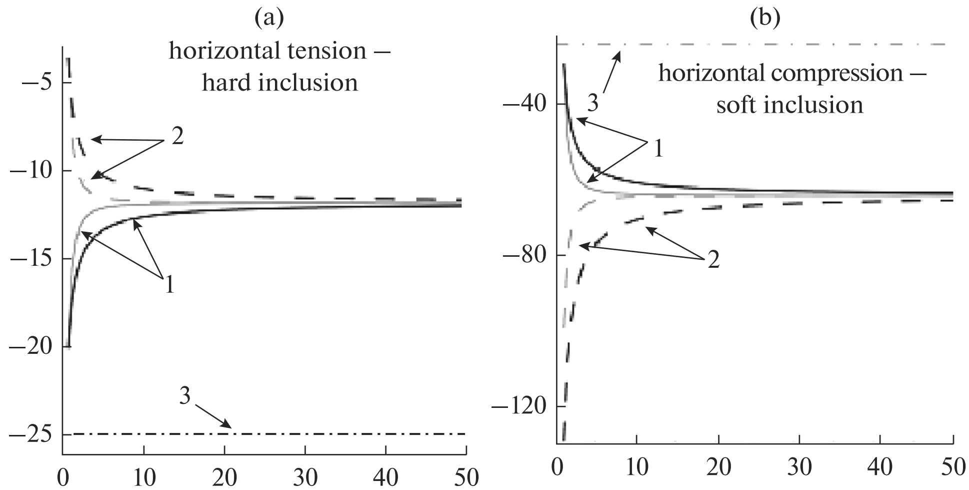

Fig. 2. The decay law for disturbances $(R/r)^n$ in radial σ_{rr} and tangential $\sigma_{\varphi\varphi}$ stresses ($\sigma_{zz} = -p_{lt} = -25$ MPa) around the inclusion in the layer. (a) the elastoplastic problem, state **Ia:** $p_o = 20$ MPa for $qp_{lt} = 11.7$ MPa (a hard inclusion), $\sigma_{\varphi\varphi} > \sigma_{rr} > \sigma_{zz}$ (horizontal tension), $n = 1$; (b) the elastoplastic problem, state **IIa** $p_0 = 30$ MPa for $qp_{lt} = 65$ MPa (a soft inclusion), $\sigma_{zz} > \sigma_{rr} > \sigma_{\varphi\varphi}$ (horizontal compression), $n = 1$. (a) and (b) also show the stresses proper to the elastic solution $n = 2$. (1, 2) radial σ_{rr} and tangential $\sigma_{\varphi\varphi}$ stresses, respectively, (3) vertical stress σ_{zz}. The solid line is for the elastoplastic solution and the dashed line for the elastic. The distance r is plotted along the horizontal axis and the stress in MPa along the vertical axis.

to make $\sigma_{\varphi\phi} > \sigma_{rr} > \sigma_{zz}$ $(\sigma_{\varphi\varphi} = \sigma_1, \sigma_{rr} = \sigma_2)$ around the inclusion. This relationship between the normal stresses that act in the horizontal plane and the upward vertical stress corresponds to the geodynamic type of horizontal tension, that which also existed during the initial phase. For this state of stress the maximum shear stress and the normal stress in the plane where it acts will be given by

$$\tau = 0.5(\sigma_{\varphi\varphi} - \sigma_{zz}), \quad \sigma_\tau = 0.5(\sigma_{\varphi\varphi} + \sigma_{zz}). \qquad (9)$$

Using (4), (9), we find that the tangential stress is constant and is equal to its value for the initial state of stress due to gravity forces alone (5):

$$\sigma_{\varphi\varphi} = -qp_{lt}, \qquad (10)$$

where q is the coefficient of lateral resistance according to (5), i.e., the initial state of horizontal tension.

Further, using the differential equation (7) and expression (10), we find:

$$\sigma_{rr} = -qp_{lt} + (qp_{lt} - p_0)\frac{R}{r}. \qquad (11)$$

The constant for this differential equation was found using the requirement at the boundary of the inclusion $(\sigma_{rr} = -p_0)$.

The radial stress tends to the limit $\sigma_{rr} \to \sigma_{\varphi\varphi} = -qp_{lt} = -11.7$ MPa ($q = 0.467$) at large distances from the inclusion $(r \to \infty)$. The requirement $\sigma_{\varphi\varphi} > \sigma_{rr} > \sigma_{zz}$ is satisfied when $p_0 > qp_{lt}$. Since this pressure exceeds the level of middle horizontal stress in the initial state, the inclusion can be treated as *hard* in the problem at hand. Figure 2 shows the variation in radial stresses for the problem of elastoplastic deformation with the same parameters as when finding the initial state of stress and for $p_0 = 20$ MPa. It follows from the resulting solution that the radial stresses attain their maximum compressive values at the inclusion contour $(\sigma_{rr} = -p_0)$ and then slowly decrease as $1/r$ toward the periphery down to $\sigma_{rr} = -qp_{lt}$.

This law of decay of the disturbed state is substantially slower than that in the elastic solution. The constant level of tangential stress is maintained by pseudoplastic (cataclastic) flow along the lines of sliding that lie on the cylindrical surface $zO\varphi$.

State IIa: horizontal compression, a soft inclusion. Let the level of pressure in the inclusion be such that we have $\sigma_{zz} > \sigma_{rr} > \sigma_{\varphi\varphi}$ $(\sigma_{\varphi\varphi} = \sigma_3, \sigma_{rr} = \sigma_2)$ around the inclusion; then this state is consistent with

$$\tau = 0.5(\sigma_{zz} - \sigma_{\varphi\varphi}), \ \sigma_\tau = 0.5(\sigma_{zz} + \sigma_{\varphi\varphi}). \qquad (12)$$

Using (4), (12), we find that the tangential stress is constant and equals its value for an initial purely gravitational state of stress (6):

$$\sigma_{\varphi\varphi} = -qp_{lt}, \qquad (13)$$

where the lateral resistance q is determined in accordance with (6), i.e., an initial horizontal compression.

Further, we use the differential equation (7) and (13) to obtain

$$\sigma_{rr} = -qp_{lt} + (qp_{lt} - p_0)\frac{R}{r}. \quad (14)$$

We also used the requirement at the boundary of the inclusion $(\sigma_{rr} = -p_0)$ to obtain (14).

At large distances from the inclusion $(r \to \infty)$ the radial stress tends to the limit $\sigma_{rr} \to \sigma_{\varphi\varphi} = -qp_{lt} = -65$ MPa. The requirement $\sigma_{zz} > \sigma_{rr} > \sigma_{\varphi\varphi}$ is satisfied when $p_0 < qp_{lt}$, i.e., the inclusion can be treated as a *soft* inclusion. Figure 2b shows the variation of radial stresses in the problem of elastoplastic deformation for the same parameters as when finding the initial state of stress and for $p_0 = 30$ MPa. It follows from the solution (13), (14) that the radial stresses have the minimum compressive values at the inclusion contour $(\sigma_{rr} = -p_0)$; afterwards they follow the law $1/r$ to decrease to $\sigma_{rr} = -qp_{lt}$ at infinity from the inclusion. This law of decay for the disturbed state is substantially slower than that in the elastic case.

The above analysis showed that elastic release occurs around the well in two cases: (1) horizontal tension and a soft inclusion (**Ib**); and (2) horizontal compression and a hard inclusion (**IIb**). The law of stress variation is then given by (8), that is, the decay of the disturbed state occurs as $1/r^2$. These two cases are of no interest to us, when we are concerned with the increase of the long-range influence.

It can also be shown that the two remaining variants of horizontal shear with a soft and a hard inclusion, (**IIIa** and **IIIb**), take place under initial horizontal compression or tension and under definite relationships between the pressure in the inclusion and the stresses at infinity. Such states of stress can arise at once near, or at some distance from, an inclusion, and afterwards become horizontal compression or horizontal tension. While we are concerned with the increase of long-range influence, neither of these two cases are of interest to us.

We conclude this section by noting that our analysis of the supercritical behavior around an inclusion relied on the Coulomb–Mohr criterion of failure; this enabled us to derive simple analytical solutions. The criterion accurately describes the behavior of cracked rock massifs in which irreversible deformations are produced, not by intracrystalline flow (true plasticity), but by cracking flow, which is cataclastic plasticity. Flow with the Coulomb–Mohr criterion is suitable for media in which crack orientations vary in short ranges relative to the shear plane. There are also other criteria that determine the phase of supercritical deformation in a cracked geological material. As an example, the Drucker–Prager criterion (Drucker and Prager, 1952) provides a better description of flow due to fine cracking (quasi-plastic flow), when crack orientations vary widely. We note that the result derived in the present study does not depend on the choice of a model for supercritical behavior, being determined by the very fact of a supercritical state during the phase that precedes the formation of the anomalous inclusion.

<u>Deep stratification in the crust following the geodynamic type of the state of stress.</u> Data from in-situ surveys show that in those crustal volumes where the maximum compressive stresses in upper horizons act in horizontal directions, their values can exceed the vertical stresses close to lithostatic pressure by multiple times (Markov, 1977, 1985; Rebetsky et al., 2017). The stress level decreases with increasing depth, tending to 0.8–0.9 of the lithostatic level. It has been found that the geodynamic type of stress in the crust is not the same over large areas and can vary with depth (Gol'din, 2004; Rebetsky and Mikhailova, 2011). Measurements of stress by in situ techniques, as is done in mining science, show that even in those crustal volumes where horizontal compression is observed in the upper parts (depths of 1.5–2 km), deeper horizons may show a change in the type of stress to horizontal shear or horizontal tension (Brady and Brown, 2004). Considered in the light of the above result, this change of stress state with increasing depth means that the long-range influence due to an incipient earthquake source *can occur differently in different layers at depth*.

On the other hand, it is known (Nikolaevsky and Sharov, 1985) that there is a seismic waveguide at depth within the crust and an increased fluid pressure is thought to exist in the waveguide. It follows that the highest probability of pseudoplastic flow is in this layer. Figure 3a shows a depth-dependent distribution in the two-layered crust for different orientations of principal stress axes relative to the vertical axis corresponding to in situ measurements as conducted by mining engineers in uplifted areas. If the upper crustal layer, which is frequently found to be under horizontal compression, according to (Brady and Brown, 2004), has not reached a supercritical state, then disturbed stresses will decay in that layer as $(R/r)^2$, as follows from the elastic solution. If we suppose that the lower layer has reached the Coulomb–Mohr limit (a waveguide) before an inclusion has been formed, then the state in that layer would correspond to our case **IIa**. The stresses that have been disturbed by the inclusion would decay as (R/r) in that layer. That means that the lower layer would serve as a conductor of the long-range influence from a hard inclusion.

If there is a soft inclusion in the two-layered structure considered above, then, as is shown in Fig. 3b, the decay law would be $(R/r)^2$ in both layers, because elastic release is occurring around the inclusion.

In this way we can observe different types of the long-range influence due to an inclusion in different layers at different depths, depending on their initial

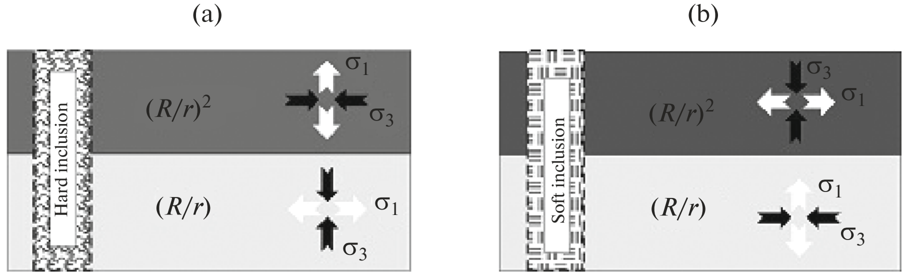

Fig. 3. Schemes illustrating the depth-dependent stratification in the massif according to the geodynamic regime of the state of stress and the inclusion type: (a) a hard inclusion, (b) a soft inclusion.

states and the relationship between the initial parameters and the parameters of the inclusion.

We also wish to note that if we interpret the above example with respect to a magma-conducting conduit, then a soft inclusion would occur during a relaxation phase when the pressure in the conduit is decreasing, while a hard inclusion corresponds to a phase of activation, when the pressure in the conduit is increasing. Hence, a nonlinear relationship is inferred between the processes in the inclusion (a magma-conducting conduit) and the surrounding rocks, which is not observed in the Dobrovol'sky elastic model.

The influence of lateral zonality in the crust. The upper crust (Sadovsky et al., 1986) can be regarded as a blocky medium with suture zones in the form of faults (Sherman et al., 1983) (Fig. 4). Large blocks in the upper crust can probably be treated as sufficiently strong features that exhibit a pure elastic response to disturbances that arise during the formation of an earthquake precursory region. In these crustal regions the rate of decay for lateral disturbances due to an anomalous zone is fairly rapid, on the order of $(R/r)^2$. On the other hand, regional and megaregional fault zones are sufficiently broad areas of defragmented rock with altered physical properties (tectonites) (Patalakha et al., 1987; Chikov, 1990, 1992, 2011). These zones should be considered at the outset as volumes in a supercritical state. Accordingly, it is in these zones that a slow lateral decay could occur according to the law (R/r).

The lateral inhomogeneity of upper crust that we pointed out above shows that the laws of stress variation around an anomalous inclusion that we have derived should be regarded as crude approximations. In fact, each concrete case requires solving its own problem for elastic blocks connected via zones of elastoplastic behavior (faults).

DISCUSSION

The chief conclusion to be drawn from the results of the study reported in this paper is that the disturbing influence of an anomalous inclusion in a material beyond the yield limit decays slower than is the case for a purely elastic model. According to I.P. Dobrovol'sky's formula, the level of strain is lower by a factor of 10^6 at a distance of $r \approx 100R$. The expressions derived in the present study, which incorporate a supercritical behavior of rocks under horizontal compression or tension, tells us that the strain level decays by 100 times only at the same distance. This is 10^4 below what would be the case for an elastic model of the material.

Latynina (2007) presented results from strainmeter observations at the Baksan station in the northwestern Caucasus and at Protvino (near Moscow) before and after the disastrous Sumatra–Andaman earthquake 2004 ($M = 9.1$). A change in the sign of strain rate was recorded at Baksan 20 days before the earthquake. The strain change was 2×10^{-7} during 10 days (a train rate of

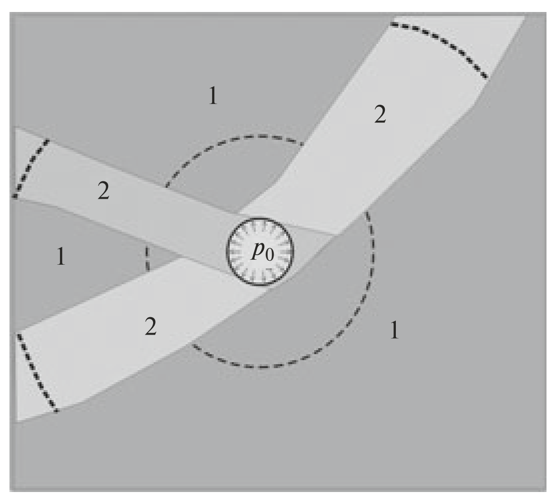

Fig. 4. A scheme illustrating the lateral fault zones and blocks structure of upper crust and differences in the long-range influence due to the precursory process in a future rupture region in fault zones (shorter thicker dashes) and in elastic blocks (plain dashed line). The same level of disturbed stresses is observed at different distances. (1) elastic blocks $(R/r)^2$, (2) fault zones in a supercritical state (R/r). The ratio between the respective dimensions is not to scale.

2.3×10^{12} sec^{-1} or about 10^{-4} yr^{-1}). Smaller amplitudes of shortening (0.8×10^{-7}) were also obtained at Protvino.

We now are going to analyze whether it could be possible to explain the deformation values measured at the two stations, having in mind the long-range influence due to the incipient rupture region of the disastrous Sumatra–Andaman earthquake. According to the results of a seismological analysis (Lay et al., 2005), the initial region of the rupture zone 420–450 km long radiated most of the seismic energy released (90–95%). It showed the highest rupture velocities (up to 3 km/s) and largest slip (up to 20 m). This patch was ascribed a magnitude $M_W \approx 8.9$. The earthquake involved a drop of shear stresses that acted along the rupture plane. Seismological estimates give the drop as equal to 0.9 MPa (Rebetsky and Marinin, 2006a, b). Assuming the shear modulus of the rock massifs to be 10^4 MPa, one can estimate the released elastic shear strain to be $\gamma \approx 10^{-4}$. The release of such a large strain led to a stabilization of the crust. It is likely that if this released elastic strain can be recovered after some lapse of time, then the medium would again become unstable and another earthquake would occur. Suppose a sharp increase in elastic shear strain occurred 10 days before an earthquake, that is, 1% of the total elastic shear strain (a strain rate of approximately 4×10^{-5} yr^{-1}; rates like these are observed in fault zones (Karmaleeva, 2011)). This strain increase is equivalent to $\varepsilon_a = \alpha\gamma = 10^{-6}$ with $\alpha = 0.01$ for the model of an anomalous inclusion.

Now, we assume the size of the earthquake source volume that has radiated the bulk of seismic energy to be the double radius of the inclusion volume $R \approx 220$ km. The distance between that patch of the Sumatra–Andaman earthquake and the Baksan station is $R \approx 8000$ km. In this case the Dobrovol'sky model would yield the strain change at Baksan equal to $\varepsilon_d = 0.01\gamma_0 (R/r)^2 = 7.5 \times 10^{-10}$ (we used (12) as being relevant to a two-dimensional problem in elasticity). If we use the approach proposed in the present paper and assume the crust to contain regions in a supercritical state in keeping with the model of an anomalous inclusion in a medium whose elastoplastic deformation is at the initial phase, then we get $\varepsilon_d = 0.01\gamma_0 (R/r) \approx 3 \times 10^{-8}$. The resulting value is in better agreement with the observations (Latynina et al., 2007) than the long-range influence as predicted by the Dobrovol'sky model.

It can be seen from the above discussion that the increase in elastic strain as fitted here for the incipient rupture zone of the Sumatra–Andaman earthquake, $\varepsilon_a = 10^{-6}$, is in good agreement with the model of elastoplastic deformation and is at variance with the model of elastic deformation around an inclusion. We should have assumed ε_a to be three orders of magnitude higher, i.e., $\varepsilon_a = 10^{-3}$ using the model of elastic deformation in order to get the strain level as observed at the Baksan station. However, this is an enormous strain, which would have given a change of 10 MPa during 10 days in the level of deviatoric stresses in the precursory region of the earthquake. Such stresses did not take place, according to estimates of stress drop for the rupture zone of the Sumatra–Andaman earthquake (Rebetsky and Marinin, 2006).

CONCLUSIONS

The interpretation of our model calculations for the actual geological medium where regions of different strength occur shows that the leading conductors of an anomalous state in the rupture zone of a future earthquake are fault zones, as well as crustal layers at depth (waveguides) where the medium is in a supercritical state. For this reason the superhigh values of the long-range influence due to an anomaly of stress that we obtained (compared with the existing standard notions) can only be observed when an anomaly and a site of observation have communication via a connected fault system or a waveguide at depth in a supercritical state.

The model we have discussed involves the decay of stresses disturbed by an inclusion following the law (R/r); the decay type is relevant to different types of inclusion (soft or hard), with the initial state being in the form of horizontal compression or tension. The *horizontal gradient* of the level of disturbed stresses in models that incorporate a supercritical state of the material is considerably below that in the Dobrovol'sky model of an elastic inclusion. The changes in the state of stress considered here for a vicinity of a cylindrical inclusion can be used to estimate the long-range influence due to an incipient earthquake. We note that this study used the Coulomb–Mohr criterion for plasticity, which disregards the form of the stress tensor (the second principal stress). Rebetsky and Lermontova (2016) showed that a simple modification of the criterion expression by inserting the mean pressure (the middle pressure taken with the opposite sign) makes the exponent n in $(R/r)^n$ dependent on the friction constant k_f. In this particular case the degree n can be diminished to reach 0.5.

The result obtained in this study can also be used for interpreting the activation of magma conduits. In that case our formulation of the geomechanical problem as stated in this paper for a vertical cylindrical cavity under pressure is even more correct. From the resulting solution it follows that when a conduit has its characteristic transverse size equal to 0.1 km and the exponent is $n = 1$, then we find that the influence of the change in magma pressure by 10 bars (a quiet effusion of magma) relative to the initial stress level can be detected in tidal strain (10^{-8}) at a distance of 300 km. Assuming $n = 0.5$ (Rebetsky and Lermontova, 2016), we find that the activation of a volcanic conduit can be detected at any distance on the Earth's surface. According to the Dobrovol'sky theory ($n = 2$), such a low level of disturbed strain must be observed at a distance of 53 km.

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