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Fractal properties of medium and seismoelectric phenomena

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Abstract

Electrokinetic phenomena in a water-porous medium with a fractal structure above percolation threshold are theoretically investigated. Fracture zone with space-variable porosity is considered as a model of an earthquake hypocenter zone in which the electrokinetic current results from fluid filtration in a fractal pore network. A critical exponent of the streaming potential coefficient is found to depend on both the transport critical exponent and correlation length critical exponent. In this model, logarithmic dependence of electric field amplitude E on the earthquake magnitude M is derived which is compatible with the one observed by the VAN group. Without fractal properties, this form of dependence contradicts the empirical data. The electromagnetic field far from the hypocenter is calculated, which leads to the prediction of weak magnetic field variations. To explain the observed amplitude of VAN's Seismic Electric Signals (SES), the electric source must be at a distance of about 10 km from the registration point if the medium is homogeneous. Therefore, some conductive channel(s) are needed to explain the long distance selective SES transmission.

1. Introduction

Electromagnetic earthquake precursors have been proposed by a number of investigators. Ultra low frequency electromagnetic noise in the atmosphere, variations of ground electric potential and other known phenomena are found to take place before earthquake occurrences (Varotsos and Alexopoulos, 1984a,b; Fraser-Smith et al., 1990; Kopytenko et al., 1990; Dea et al., 1991; Johnston et al., 1994; Hayakawa et al., 1996, 2000; Uyeda et al., 2000). The so-called VAN-method of measuring Seismic Electric Signals (SES) at some days or weeks before earthquake occurrences has been used in Greece for earthquake forecasting for more than 15 years. The

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remarkable property of SES is that it can be recorded at sensitive sites which are a hundred or more kilometers from the epicenter. Moreover, a sensitive site is sensitive only to SES from some specific focal area(s). These properties called "Selectivity" can not be explained by a homogeneous medium (Varotsos et al., 1996, 1998).

The empirical dependence of SES amplitude E (μV/m) on earthquake magnitude M looks as (Varotsos et al., 1996).

$$\log E = aM + b, \quad a \approx 0.34 - 0.37 \tag{1}$$

where a and b are empirical constants. The value of b depends on the azimuth of epicenter reckoned from observation station and the "sensitivity" of station. In other words, the parameter b is not universal.

The recent discussion on the VAN-method has divided the scientific community into two: one supporting it and the other rejecting (e.g. Jackson and Kagan, 1998; Uyeda, 1998). A number of attempts have been made to give a theoretical explanation of SES. Hypotheses of piezo-stimulated current and current generated by charged dislocations have been proposed by Varotsos and Alexopoulos (1986) and Slifkin (1993). Some theories are based on the electrokinetic hypothesis (Dobrovolsky et al., 1989).

The electrokinetic currents can be observed in water-saturated media with fluid-filled channels (Mizutani and Ishido, 1976; Jouniaux and Pozzi, 1999). The walls of pores and cracks in a solid body generally adsorb cations from the liquid. Moving along the channel, the liquid carries ions of opposite sign, and thus produces an extrinsic electric current. The electrokinetic current density averaged with respect to the cross section can be written as

$$\mathbf{j}_e = -C_e \nabla P, \tag{2}$$

$$C_e = \sigma C \sim \frac{\varepsilon \varepsilon_0 \zeta n}{\eta}, \tag{3}$$

where C is streaming potential coefficient, σ is rock conductivity, ∇P is gradient of pore pressure, ε and η are dielectric permeability and viscosity of fluid, ε_0 is electric constant, ζ denotes the potential difference across the electric double layer on pore walls (ζ-potential) and n is medium porosity. Notice that P is the excess fluid pressure over the hydrostatic pressure.

If the source of electrokinetic currents is placed in the hypocentral zone, the SES amplitude would be much less than the observed one because of rapid attenuation of currents with distance (Bernard, 1992). However, Varotsos et al. (1996, 1998) have demonstrated theoretically that if there are electrically highly conductive channels between focal zone and SES sensitive sites, selective transmission of SES over distances of hundreds kilometers is possible, especially when the conductive path ends near the observation site. Actually, the theory predicts that SES sensitive sites are expected to be in the areas with sizes of a few km directly above the hypocenter and above the end of the channel.

The main goal of this paper is a theoretical analysis of the influence of possible fractal structure in earthquake hypocenter zone on seismoelectric phenomena.

2. Fractal critical exponents of electrokinetic current parameters

Suppose that an earthquake hypocenter is surrounded by water-saturated porous rocks with fluid-filled pore channels. The pre-earthquake stage is accompanied by appearances of a number of fresh cracks in the vicinity of hypocenter. We call such a zone a fracture zone. The scale of the fracture zone may be varied from hundreds of meters up to several kilometers. We assume that the pore space in the fracture zone exhibits fractal structure (Feder, 1988; Iudin et al., 1999). Apparently, most of the fresh cracks are closed when formed. Because of the pressure release due to cracking, they are under lower pressure, so that water from uncracked outer region can penetrate into them as soon as a network of connected channels or fractal clusters is formed. The closed fresh cracks may be regarded as the sink of water from surrounding higher pressure areas.

Let us suppose the porosity n and permeability of rocks, after the cluster formation, decreases from the center of the fracture zone towards the periphery by a certain law, as shown in Fig. 1. The percolation threshold n_c is exceeded in the internal area with typical size L. It means that the permeability tends to zero outside this zone. Actually, there is a finite permeability due to the fact that crustal rocks contain a wide range of small cracks that can be connected. Further we are interested in conductivity of the rock rather than its permeability. The conductivity of the surrounding space is also non-zero due to both the bulk and surface conductivities of the small fluid-filled cracks and conductivity of the rocks itself. We suppose that these conductivities can be neglected in comparison with that of the fluid-filled cracks, which are formed in the fracture zone, i.e. the conductivity outside the fracture zone is negligible. It means that the value n_c is rather related to the percolation threshold for conductivity due to the fresh fluid-filled cracks. It should be emphasized that a variety of the crack sizes can be described only in the framework of rather complicated percolation theory. Below we restrict our analysis by a simple percolation theory without of account of the crack/channel size distribution. Then the fractal properties near the threshold are determined by the correlation length (ζ).

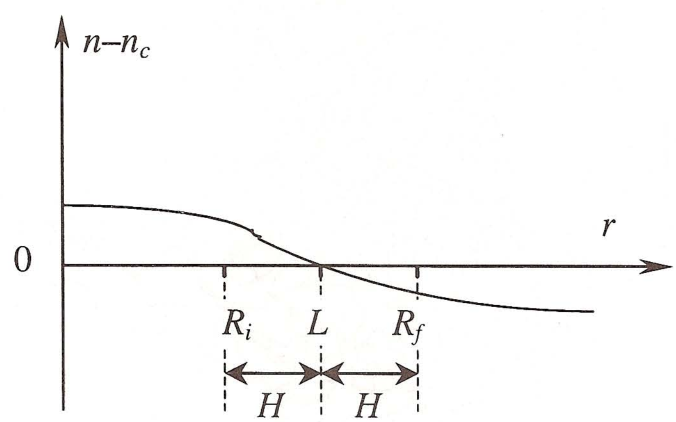

Fig. 1. Assumed approximate dependence of the porosity n on distance r in the fracture zone. The percolation threshold n_c is exceeded in the internal area $r < L$, and non-fractal central zone is situated at $r < R_i$. The fractal region is confined by the radii R_i and R_f.

$$\xi \sim \frac{1}{|p - p_c|^{\nu}} \sim \frac{1}{|n - n_c|^{\nu}}, \tag{4}$$

where $\nu = 0.88$ is the correlation length critical exponent, p is probability that a channel can conduct the fluid, and p_c denotes the critical probability related to percolation threshold (Feder, 1988; Staüffer, 1979). ξ represents the typical size of the fractal cluster. According to Eq. (4), an increase in porosity in the internal area $(r < L)$ is followed by a decrease in correlation length. So, the central part of the internal area can lose the fractal properties because of multiple intersection of channels. This part with high porosity and high permeability will be called non-fractal central zone in contrast to the peripheral fractal region. Let $R_i = L - H$ be the typical radius of the non-fractal central zone as shown in Figs. 1 and 2. The fractal region occupies a spherical shell confined by the radii R_i and $R_f = L + H$. It is obvious that typical size H of the fractal region would be of the order of the correlation length (4), i.e. $H \sim \xi(R_i) \sim \xi(R_f)$. Further we represent the porosity near $r = L$ as a power series of parameter H:

$$n(r) \approx n_c + \frac{dn}{dr} H. \tag{5}$$

Let us substitute the expression (5) into Eq. (4). Using the rough approximation $dn/dr \sim \Delta n/L$ at $r = L$, where Δn is typically the difference of n at $r = 0$ and $r = L$ we can find the dependence of H on the size of the internal area L:

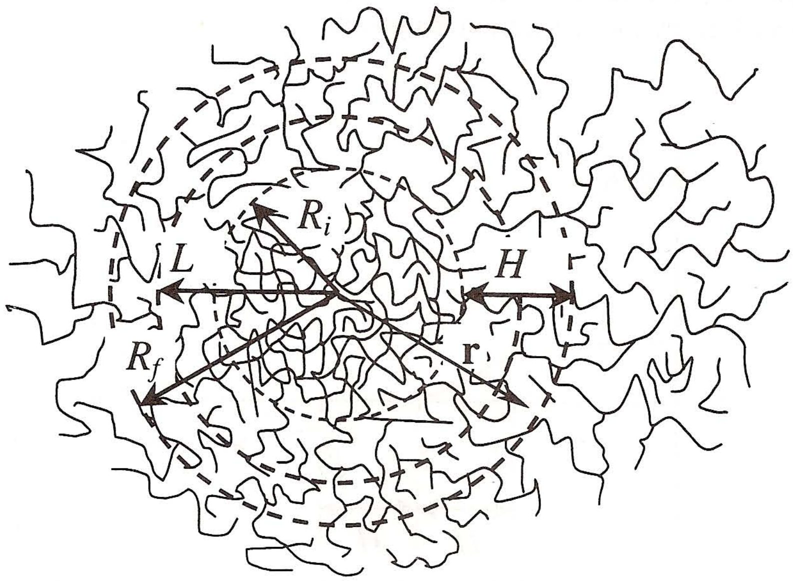

Fig. 2. Schematic picture of the fracture zone. Internal high permeability area is restricted by the radius R_i. The fractal region occupies the field from $r = R_i$ to $r = R_f = R_i + H$. L denotes the distance where $n = n_c$.

$$H \sim \left(\frac{\mathrm{d}n}{\mathrm{d}r}\right)^{\left(-\frac{\nu}{\nu+1}\right)} \sim L^{\nu/(\nu+1)}. \tag{6}$$

The expression (6) is valid if $H << L$. For this case we can suppose that the size of the fracture zone is of the same order as that of the internal area, i.e. L, and so the dependence of H on the fracture zone size looks like Eq. (6). We need this H dependence on fracture zone for our later discussion on SES dependence on earthquake magnitude.

Suppose that porosity varies weakly in the central non-fractal high-permeability zone (Fig. 1), so that the electrokinetic coefficient C_e can be considered as constant. Otherwise let us replace coefficient C_e by its averaged value. But in the fractal peripheral region, C_e as well as diffusion coefficient, permeability and other rock parameters, vary with distance by power law (Stauffer, 1979; Feder, 1988). We chose the dependence of C_e on distance in the following form:

$$C_{\mathrm{e}} \sim \frac{G}{r^{\mu}}. \tag{7}$$

Here G is a constant, μ is unknown streaming potential critical exponent and r is the distance between two arbitrary points in the fractal region. Now, let us find the value of μ.

If the porosity is close to the percolation threshold, the coefficient C_{e} in the form (3) is not applied because the fractal geometry on n should be taken into account. For a separate channel, the electrokinetic current density $\mathbf{j}_{\mathrm{e}}^{(1)}$ is described by the known equation

$$\mathbf{j}_{\mathrm{e}}^{(1)}\mathbf{E}_{\mathrm{eff}}, \tag{8}$$

$$\mathbf{E}_{\mathrm{eff}} = -\frac{\varepsilon\varepsilon_0\zeta}{\eta\sigma_{\mathrm{f}}}\nabla P. \tag{9}$$

Here $\mathbf{E}_{\mathrm{eff}}$ denotes the effective field strength of extrinsic forces and σ_{f} is the fluid conductivity. For a fractal medium, we need to average Eq. (8) over its cross section. Instead of this procedure, let us use the following expression for the average current density:

$$\mathbf{j}_{\mathrm{e}} = \sigma_{\mathrm{r}}\mathbf{E}_{\mathrm{eff}} = -C_{\mathrm{e}}\nabla P, \tag{10}$$

$$C_{\mathrm{e}} = \frac{\varepsilon\varepsilon_0\zeta\sigma_{\mathrm{r}}}{\eta\sigma_{\mathrm{f}}}. \tag{11}$$

where σ_{r} denotes average conductivity of the rock containing pore channels. As results from the percolation theory, the rock conductivity σ_{r} depends on the porosity as follows:

$$\sigma_{\mathrm{r}} \sim (n - n_{\mathrm{c}})^{t}, \tag{12}$$

where $t \approx 1.6$ is the transport critical exponent (Stauffer, 1979; Feder, 1988). This value of t, as well as other critical exponents, is derived by means of numerical simulation on three-dimensional grids. So, on account of Eqs. (11) and (12), we can conclude that

$$C_e \sim \sigma_r \sim (n - n_c)^t. \tag{13}$$

The fluid can move only along the channels belonging to fractal clusters. We introduce active porosity n_a: the volumetric part of the rocks occupied by the percolation clusters, which can be expressed through usual porosity n

$$n_a \sim (n - n_c)^\beta. \tag{14}$$

Here β is the order parameter critical exponent that can be expressed in the form: $\beta = (3 - D_f)v$, where $D_f = 2.54$ denotes fractal dimension of the percolation cluster (Stauffer, 1979; Feder, 1988). The active porosity varies in the fractal region as $r^{D_f - 3}$, where r s the distance between two arbitrary points in the fractal region. So, we can estimate this value as follows: $n_a \sim H^{D_f - 3}$. Using Eqs. (13) and (14) and the above estimation of n_a, one can find $C_e \sim n_a^{t/\beta} \sim H^{-t/v}$. Comparing this expression with Eq. (7), we finally get

$$\mu = t/v \approx 1.82 \tag{15}$$

At far distance from the fracture zone, the low-frequency electromagnetic field generated by a system of electrokinetic currents can be characterized by an effective short linear current element $I\Delta\mathbf{l}$. Here I denotes the total source current and $\Delta\mathbf{l}$ is the effective current length. To replace an extrinsic current system by a point current element, it is necessary to integrate the electrokinetic current density over the volume V of the fracture zone. Using Eq. (2) we get:

$$\mathbf{p} = I\Delta\mathbf{l} = \int_V \mathbf{j}_e \mathrm{d}V = -\int_V C_e \nabla P \mathrm{d}V. \tag{16}$$

Let us find the dependence of value $\mathbf{p}$ on fracture zone size L. The contribution to the integral (16) of the central high-permeability zone with constant electrokinetic coefficient C_{e0} is approximately proportional to the value $p_c = C_{e0}|\nabla P|L^3$. The simple estimation of pore fluid pressure gradient is $|\nabla P| \sim \Delta P/L$, where ΔP is the pore pressure difference between high- and low-pressure regions. Taking into account that shear stress drop caused by micro-fracturing process is independent of the size L (Scholz, 1990) we assume the same as to ΔP, i.e. ΔP is independent of the size L. Thus, we get $p_e \sim L^2$. The last estimation is not valid if the contribution of the central high-permeability zone is negligible, for example, if the electrokinetic current distribution is approximately spherically symmetric in this zone. In such a case, the contribution of this zone to the integral (16) can be negligibly small because we add the symmetrically distributed vectorial values.

The contribution of p_f fractal region to the integral can be similarly estimated if we take into account the expression (6) and dependence (7) as,

$$C_e \sim H^{-\mu} \sim L^{-t/(1+v)}. \tag{17}$$

On account of estimation (17) one can find

$$p_{\mathrm{f}} \sim C_{\mathrm{e}} \frac{\Delta P}{L} L^2 H \sim L^{1-(t-v)/(1+v)} \tag{18}$$

If SES results from the electrokinetic current developed in the fracture region, SES amplitude E must be proportional to the amplitude of $\mathbf{p}$. Let us find the relationship of E and the magnitude M of earthquake in preparation. The dependence of earthquake focus dimension L on the earthquake magnitude M can be described by the known empirical equation (Kanamon and Anderson, 1975):

$$\log L = 0.5M - 1.9 \tag{19}$$

If the main contribution to the total electric field comes from the central non-fractal zone, we get $E \sim p_{\mathrm{c}} \sim L^2$ [cf. the paragraph below Eq. (16)]. Putting this relation into Eq. (19), we find the following dependence: $\log E = \mathrm{M} + \mathrm{constant}$, which is obviously in contradiction to the empirical relation (1).

If, on the other hand, we assume that the contribution of fractal region is much larger than that of the central zone, i.e. $p_{\mathrm{f}} >> p_{\mathrm{c}}$, we obtain a relation compatible to experimental data as follows. In this case $E \sim p_{\mathrm{f}}$. Taking into account Eq. (18), one can find that Eq. (19) gives,

$$\log E = aM + b. \qquad a = 0.5\left(1 - \frac{1-v}{1+v}\right) \approx 0.31 \tag{20}$$

Here we used the following critical exponents: $t = 1.6$, $v = 0.88$ obtained by numerical simulation on three-dimensional grids (Stauffer, 1979; Feder, 1988). The value of parameter a in Eq. (20) is very close to the observational one (1). Fractal structure in the fracture region is needed to explain the experimental rule (1). This means that the contribution from the central non-fractal zone must be small compared to that from the fractal zone, which in turn requires, for example, that current distribution in the central zone is approximately spherically symmetric, whereas it is asymmetric in the fractal region.

3. Estimation of electric field at far distance

At a distance $r >> L$ far from the earthquake epicenter, the current source may be considered as a point. Let us consider the ground as a conductive half-space $z < 0$ with constant conductivity σ. The upper half-space $z > 0$ simulating the atmosphere is an insulator. The current element $\mathbf{p} = I\Delta\mathbf{l}$ with arbitrary direction is situated at the depth $z = -h$ on z axis, and the chosen coordinate system is shown in Fig. 3. The vector $\mathbf{p}$, which can be time dependent, is placed in the x–z plane, so that the vector $\mathbf{p}$ has only two components: p_z and p_x. If the skin-length is much larger than the typical distances, a quasi-static approximation is valid and the variations of electric field $\mathbf{E}$ and magnetic field $\mathbf{B}$ at the ground surface $z = 0$ looks as

$$\varphi = \frac{p_z h + p_x r \cos\psi}{2\pi\sigma R^3}, \qquad R = \sqrt{h^2 + r^2}, \tag{21}$$

$$E_z(0+) = \frac{p_z(2h^2 - r^2) + 3p_x rh\cos\psi}{2\pi\sigma R^5}, \qquad E_z(0-) = 0 \tag{22}$$

$$E_r = \frac{3p_z rh + p_x(2r^2 - h^2)\cos\psi}{2\pi\sigma R^5} \tag{23}$$

$$E_\psi = \frac{p_x \sin\psi}{2\pi\sigma R^3} \tag{24}$$

$$B_z = \frac{\mu_0 p_x r \sin\psi}{4\pi R^3} \tag{25}$$

$$B_r = \frac{\mu_0 p_x \sin\psi}{4\pi R^2}\left(\frac{R}{h+R} - \frac{h}{R}\right), \tag{26}$$

$$B_\psi = -\frac{\mu_0 p_x \cos\psi}{4\pi R(h+R)}. \tag{27}$$

Here φ is electric potential and μ_0 is magnetic constant. The parameter φ is the angle between the vector **r** and positive direction of x-axis (Fig. 3). In this approximation the time-dependence of electromagnetic fields is determined by the functions p_z(t) and $p_x(t)$.

To estimate the amplitude of the electromagnetic fields, the following parameters were chosen: $C=10^{-6}$–10^{-8} V/Pa, $\sigma=10^{-3}$ S/m, $\Delta P\sim 1$–10 MPa, $L=10$ km, $h=20$ km, $r=100$ km. These parameters are in agreement with value of pore fluid pressure gradient $\nabla P \sim 10^2 - 10^4$ Pa/m that usually assumed for $M=6$–7 (Ishido and Mizutani, 1981). Based on Eqs. (18), (19) and (25)–(27) we get

$$E_r \sim \frac{p_f}{\pi\sigma R^3} \sim \frac{1}{\pi} C\Delta P \frac{L^2}{R^3} \sim 0.3 - 3\cdot 10^{-4} \mu\text{V/m}, \tag{28}$$

$$B \sim \frac{\mu_0 p_f}{4\pi R^2} \sim \frac{\mu_0 \sigma C \Delta P}{4\pi}\frac{L^2}{R^2} \sim 10 - 0.01 \text{ pT}. \tag{29}$$

The weakness of magnetic variations in Eq. (29) might be expected a priori because the electrokinetic current (2) is related to a curl-free current source. Notice that the magnetic variations (25)–(27) are exactly equal to zero in the case of vertical source ($p_x=0$). Thus, the theoretical conclusion on the weakness of electrokinetic current-induced magnetic variations during the SES appearance agrees with the observational data (Varotsos et al., 1996) as well as theoretical expectation (Sarlis et al., 1999).

The experimental amplitude of SES-signal was about 10 μV/m at the distance of the order of 100 km from the epicenter (Varotsos et al., 1996). The theoretical prediction for the electric field (28) is much less than the observed one unless we suppose a distance to the epicenter is as small as about 20–30 km. Therefore, one needs to assume a conductive channel in the crust for long distance transmission of SES (Varotsos et al., 1998). The effect can be also caused by a source of the electrokinetic current situated in the vicinity of the observation site. For instance, Bernard (1992)

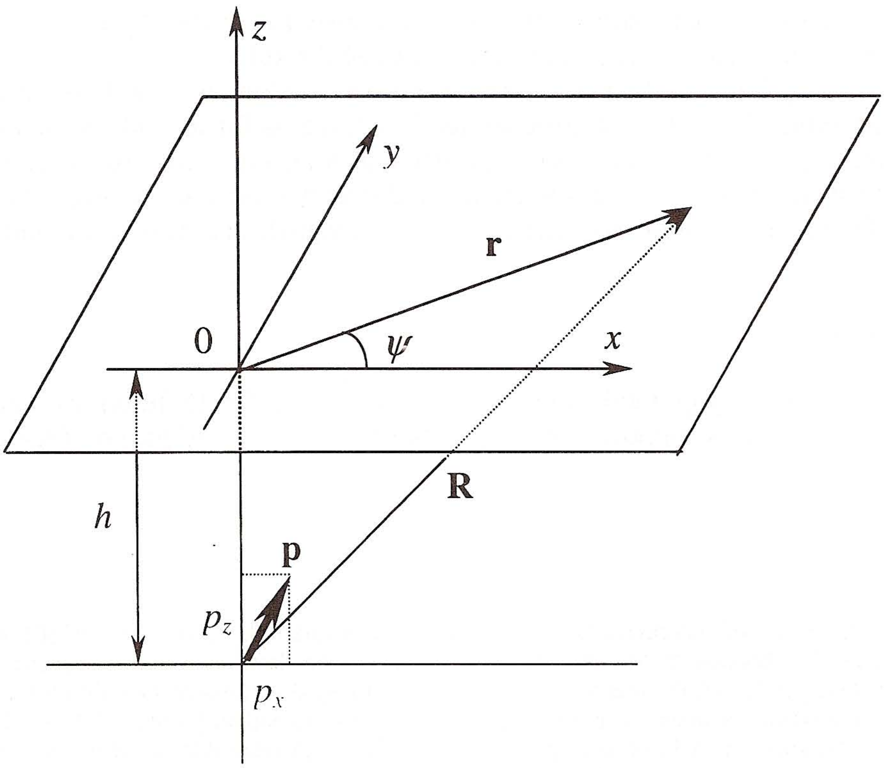

Fig. 3. The point current element $\mathbf{p}=I\Delta\mathbf{l}$ that is situated in the x–z plane in a conductive half-space $z<0$.

and Fenoglio et al. (1995) have suggested that fluid filtration from sealed high pore pressure compartments towards low pressure region may be triggered by weak seismic events to generate electrokinetic effects.

4. Conclusions

Electrokinetic effects possibly associated with earthquake preparation process have been theoretically investigated, with the assumption that structure of pore space around hypocentral zone has fractal properties. Following results have been obtained;

1. The electrokinetic coefficient C_e varies in the fractal region, and the critical exponent of the coefficient C_e is $\mu=t/v\approx1.82$, where t denotes the transport critical exponent and v denotes the correlation length critical exponent.
2. The predicted dependence of SES amplitude E on earthquake magnitude M, when fractal properties of fracture zone are assumed, is: $\log E=aM+b$, where $a=0.5(1-(t-v)/(1+v))\approx0.31$. This theoretical value of coefficient a is very close to the empirical one. If the fracture zone is not fractal, we obtain $a=1$ which contradicts with the observational data.

3. The estimations of magnetic variations at far distance caused by electrokinetic current source give very weak values, which agree with SES results.
4. Predicted amplitude of SES in a homogeneous medium is found to be less than the observed value if the electrokinetic source is located as far as 100 km or more from the observation point. To explain the experimental data, one needs to assume some highly conductive channel(s) between the source and observation sites as proposed by the VAN group. Otherwise the sources of the electrokinetic signals are local to the observation sites.

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