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Electric Current Generation Associated with the Deformation Rate of a Solid: Preseismic and Coseismic Signals

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Abstract. We present a model for the generation of electric current in rocks under stress, involving the strain rate, ($\dot{\varepsilon}$) which is influenced by the motion of charge bearing dislocations. The relationship between current density and strain rate is demonstrated. On the basis of laboratory data, we estimate the deformation rate necessary to generate an electric signal observable at distances far enough from the source, as to qualify it as an electric earthquake precursor. Using this mechanism and the geometrical characteristics of such a type of source we simulate the propagation of the electric signal and its 'received' characteristics as a function of the source-receiver separation. We conclude that the expected signal waveforms at long distances from such a kind of source are similar to a class of signals (bay like waveforms), independently observed prior to earthquakes by several investigators.

1 Introduction

The generation of electric currents in rocks under stress has long been recognized by means of experiment. For instance, Whitworth (1975) has demonstrated such an effect in alkali halides under pressure. In a more recent paper (Hadjicontis and Mavromatou, 1994), the generation of electric current in rocks under uniaxial compression is reported. The main result of these investigations is that the shape of the electric signal depends drastically on the rate of stress variation and specifically, appears to follow the first derivative of the externally applied stress. From a theoretical point of view, two types of current generation mechanisms have been studied. The first is reported by Varotsos et al. (1993) and is related to polarization or depolarization effects. The second, (Slifkin, 1993), is based on the concept of motion of charged dislocations. In the present paper, we discuss the available data on the basis of the moving charged dislocations (MCD) model. Furthermore, and assuming that the earthquake source may be described by such a model, we make a first attempt to simulate the propagation of the resulting transient electric signal and describe its received characteristics at intermediate - long distances from the source.

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2 A current generation mechanism in rocks under pressure : Geophysical implications.

In a crystalline structure, dislocations may be formed by an excess or absence of a half-plane of atoms. The edge of this half-plane comprises a dislocation line, around which the physical fields related with it are concentrated. In an ionic structure there will be an excess or absence of a line row of ions along the dislocation line, with consequence that the dislocation becomes charged. We note that the jogs of edge dislocations can be ionically charged. In a thermal equilibrium, dislocation lines are electrically neutral (Whitworth, 1975). In dynamic processes, when dislocations move, neutrality can no longer be maintained.

Let Λ^+ be the density of edge dislocations of the type required to accommodate the uniaxial compression (or tension) and Λ^- be the density of dislocation of the opposite type. The motion of charged dislocations produces a transverse polarization:

$$P = (\Lambda^+ - \Lambda^-)\cdot q_l \cdot \frac{\delta x}{\sqrt{2}} = \delta\Lambda \cdot q_l \cdot \frac{\delta x}{\sqrt{2}} \tag{1}$$

where q_l is the charge per unit length on the dislocation. If screw dislocations are ignored, the plastic contribution to the strain, when these dislocations of Burger vector **b** move through a distance δx, is

$$\varepsilon_P = (\Lambda^+ - \Lambda^-)\cdot \mathbf{b} \cdot \frac{\delta x}{2} \tag{2}$$

The rate of change in polarization is equivalent to the electric current density, by definition

$$J_P = \frac{\partial P}{\partial t} \xrightarrow{(1),(2)} J_P = \frac{\sqrt{2}}{\beta} \cdot \frac{q_l}{\mathbf{b}} \cdot \dot{\varepsilon}_P, \beta = \frac{\Lambda^+ + \Lambda^-}{\Lambda^+ - \Lambda^-} \quad (3)$$

where β is a constant usually between 1 and 1.5. Equation (3) shows that the observed transient electric variation is related to the non-stationary accumulation of deformation. Specifically, when the deformation rate increases at a variable rate (i.e., $\dot{\varepsilon} > 0$ and $\ddot{\varepsilon} \neq 0$), then : **1)** if $\ddot{\varepsilon} > 0$ ($\ddot{\varepsilon} < 0$) then $\dot{E} > 0$ ($\dot{E} < 0$) - the observed electric field variation increases (decreases); **2)** if $\dot{\varepsilon} > 0$ and $\ddot{\varepsilon} = 0$ ($\dot{\varepsilon}$ constant) then E is constant. For a number of alkali and silver halides, experimental values have been deduced for the charge density, a typical value is of the order of 0.1e/α where e is the electronic charge and α the lattice spacing (Whitworth, 1975). If we assume comparable figures for the rocks and let $\alpha \approx 5\times10^{-10}$ m, the charge per unit length is of the order of 3×10^{-11} Cb/m (Slifkin, 1993). Assuming $\dot{\sigma} \approx Y\dot{\varepsilon}$, (Y the Young modulus), the stress sensitivity coefficient F originally introduced by Nomicos and Vallianatos, (1997) is :

$$F = \frac{E}{d\sigma / dt} \approx \frac{\sqrt{2}}{\beta} \cdot \frac{q_l}{\mathbf{b}} \cdot \frac{\rho}{Y} \quad (4)$$

where E is the observed electric field and ρ is the specific resistivity of the material. We now proceed to estimate the order of magnitude of F, the stress sensitivity coefficient. Introducing in eq.(4) the values $\mathbf{b} \approx 5\times10^{-10}$m (Slifkin, 1993), $\rho \approx 10^7\Omega$m, and Y ≈ 0.8 Mbar (Vallianatos, 1996), we find $E/(d\sigma/dt) \approx 0.75$ (V/m)/(bar/s), which is comparable to the values calculated using experimental data (Table 1).

Table 1. Stress sensitivity coefficient for various rock types, estimated using data from Hadjicontis and Mavromatou (1994).

Type of rock	$E/(d\sigma/dt)$ [(V/m)/(bar/s)]
Ioannina limestone	0.70
Granite (sample A)	0.46
Granite (sample B)	0.32
mineral quartz	1.14

We shall, now, attempt now to present an analysis of the observed preseismic variations based on the MCD model (Slifkin, 1993; Nomicos and Vallianatos, 1997). According to this model all rocks contain crystalline materials which already bear linear defects (i.e., charged edge dislocations), either inherently or/and due to previous loading (deformation) cycles. Now consider that an earthquake zone comprises a volume filled by cracks or, equivalently, dislocation arrays (i.e. a concentration of charged dislocations). Thus, the processes of earthquake preparation (the non-linear evolution of stress) and energy release (rupture) are determined by the evolution, propagation and nucleation of these defects and changes in their density.

This model of the earthquake source allows for the generation of electric current with intensity proportional to the velocity of propagation of the MCD and to their density. The current density vector at the source is parallel to the velocity vector. Then, the horizontal component of the electric field, measured at a point on the surface with an epicentral distance x from the source and at time t_κ, may be qualitatively expressed as (Ernst et al, 1993):

$$E(x,t_\kappa) \approx sf_1\rho\sum(\rho_{di}\cdot \upsilon_{mi})\frac{x - x_i(t_\kappa)}{R_i^n(t_\kappa)} \quad (5a)$$

where R_i is the distance of the observation point from the MCD element located at x_i, n=2 for a linear source, ρ_{di} is the CD density at point x_i, υ_{mi} is the dislocation velocity at x_i, s is a sensitivity coefficient at the location of the receiver and f_1 is a coefficient dependent on the geometry of the source. We emphasize that expression (5a) is based on the assumption that the time sequence of stress evolution is the same at each point on the source plane, only shifted by the time delay required for crack propagation. However, the macroscopic deformation rate, $\dot{\varepsilon}_i$ is related to the MCD density and velocity via a generalization of Orowan's law ($\dot{\varepsilon} \approx \rho_{di}\cdot \upsilon_i$), and therefore (5a) becomes (Nomicos and Vallianatos, 1997):

$$E(x,t_\kappa) \approx sf_2\rho\sum\dot{\varepsilon}_i\frac{x - x_i(t_\kappa)}{R_i^n(t_\kappa)} \quad (5b)$$

where f_2 is also a factor dependent on the source. According to many models (Myachkin et al, 1986), the development of processes prior to an earthquake may be qualitatively described in terms of four stages. Consider a volume within a statistically homogeneous medium, subject to stress increasing with time. During Stage I there is a gradual increase in crack density; this Stage does not appear to associate with electric preseismic phenomena. When crack density exceeds a certain critical level, the natural interaction of cracks accelerates their development, causing an avalanche during Stage II. Thus, we observe a rapid increase in strain rate and the appearance of an anomalous electric field variation. In the last Stage IV of the process (failure) a "co-seismic" electric pulse may appear. Stuart (1988) indicated that strain accumulation should become non-linear near the end of the loading cycle. Thus considerably accelerated deformation effects should be produced during the last several days prior to the earthquake; this defines the duration of Stage II and is consistent with the observed time lag between the anomalous electric field variation and the ensuing earthquake (Nomicos and Vallianatos, 1997).

Let us, now, consider the deformation rate necessary to generate the level of the observed preseismic transient electric variations, which are of the order of 2×10^{-5} V/m. Using the stress sensitivity for the granite of Table 1, a stress rate of 6.6×10^{-5} bar/s would be required, assuming that the signal does not decay over the distance between the receiver and the hypocenter. We assume a homogeneous, linear, isotropic medium with a Young's modulus of 0.66 Mbar. Then the stress rate is converted to a strain rate of 10^{-10} s^{-1}. Ioannina limestone, with Young modulus 0.94 Mbar, yields a strain rate of 3×10^{-11} s^{-1}. Such strain rates are not impossible to achieve (S.Park, private communication). For instance, precursory strain changes are clearly illustrated in Wakita (1988), for the M7 Izu-Oshimakinkai earthquake of 14/1/1978; these indicate that strain rate could easily have reached the value of 10^{-11}-10^{-12} s^{-1}.

3 Propagation of the electric signal

The MCD current generation mechanism described in the foregoing, amounts to the superposition of a large number of dipole sources, each dipole being an individual propagating crack or a cluster of simultaneously moving dislocations. In the following, we will attempt to investigate the behaviour of this type of source at distances of several tens to a few hundreds of km. In doing so,

1. We assume that the source is a 3-D volume (cell) within the earthquake preparation zone filled with cracks which are excited coherently. Coherence in not a physical requirement but only a convenient first approximation.
2. By superimposing individual dipole fields we estimate upper limits for the expected amplitudes of the received transient fields.
3. We attempt to gain insight into the received characteristics (waveforms) of the transient fields by simulating the propagation of a pulse from an individual dipole and an individual cell. In order to demonstrate our arguments, we only consider propagation in a uniform half space.

The electromagnetic field at a point z in a conducting half space, due to a horizontal electric dipole buried at depth d, can be expressed in terms of a sum of the direct field, the ideal reflected field or field of an ideal image and the rest of the field (King et al, 1992). The latter includes a lateral-wave field at the earth-air interface and correction terms for the reflected field to account for the fact that it may be not, that of an ideal image. Due to the low frequencies involved, the lateral-wave field is insignificant and may be omitted. Inasmuch as we are interested in the behaviour of the field at intermediate-large distances from the source, we assume that the source-receiver separation is considerably greater than the depth of burial and $R \geq 5d$, $R \geq 5|z|$, so that the epicentral distance $r \sim [R^2 + (z-d)^2]^{1/2} \sim R$. At $z=0$ and after considerable algebra, we may simply express the radial electric field as:

$$E_R = -\frac{i\rho\cos\phi}{2\pi}\left[\left(\frac{1}{\delta R^2}+\frac{d}{\delta R^3}+\frac{3d}{2R^4}\right)+\ldots + i\left(\frac{1}{\delta R^2}-\frac{d}{\delta R^3}+\frac{1}{R^3}\right)\right]e^{-\frac{(1-i)R}{\delta}} \tag{6a}$$

where d is the depth of burial, δ is the skin depth and σ the conductivity of the earth. Likewise, one may derive an expression for the azimuthal electric field:

$$E_\phi = \frac{i\rho\sin\phi}{2\pi}\left[\left(\frac{1}{\delta R^2}-\frac{3d}{2\delta R^3}-\frac{5d}{8R^4}\right)+\ldots + i\left(\frac{1}{\delta R^2}+\frac{1}{R^3}+\frac{2d}{\delta^2 R^2}+\frac{3d}{2\delta R^3}\right)\right]e^{-\frac{(1-i)R}{\delta}} \tag{6b}$$

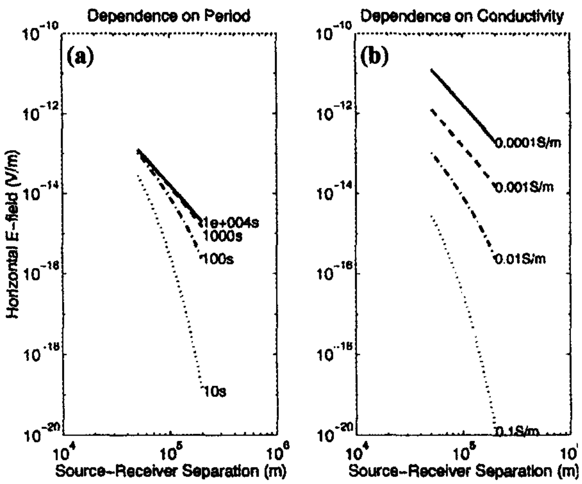

Fig. 1. Attenuation of the electric field from a unit dipole crack as a function of source-receiver separation R, for (a) different periods and, (b) different conductivities of the earth medium.

4 Propagation and waveforms of the dipole field from a single crack

First, let us examine the dependence of the received electric field on the distance from the source (source-receiver separation). The source has a unit current moment, it is buried at d=10000m and its time function is assumed to be a Kronecker-δ with an infinite spectral density of unity. The results are shown in Fig. 2, for the total electric field propagating over distances of 50-200km.

Fig. 1a illustrates the dependence as a function of frequency (period), in an Earth medium with constant σ=0.01S/m. It may easily be verified that $E(R) \sim R^{-3}$ at long periods (10^4s), changing to $E(R) \sim R^{-4}$ at shorter periods (100s). Notably, at T<100s, the dependence appears to follow a much faster decay law, which changes from approximately R^{-5} at ranges of 50-60km to the extremely fast rate of R^{-10} and higher at ranges of 150-200 km. Fig. 1b illustrates the dependence as a function of conductivity at constant period T=500s. An approximate R^{-3} law appears to control field decay, save for the more conductive half-space (0.1S/m), where a gradual transition from R^{-3} closer to the source, to R^{-6} at 150-200km appears to exist.

Such behaviour of the electric field may be understood from its dependence on skin depth. Large skin depths imply that at distances of 100-200km, we are in near field conditions and decay is controlled by the product $\delta^1 R^{-2}$. Smaller skin depths amount to departure from near field conditions and decay is gradually dominated by the exponential and R^{-2} factors. Such conditions are satisfied for combinations of large distances and high conductivities, or short periods, or both. Thus, it appears that the short period (e.g. <1s) components of the transient signal are practically impossible to propagate beyond a few tens of km, even under very favourable conditions. At large distances

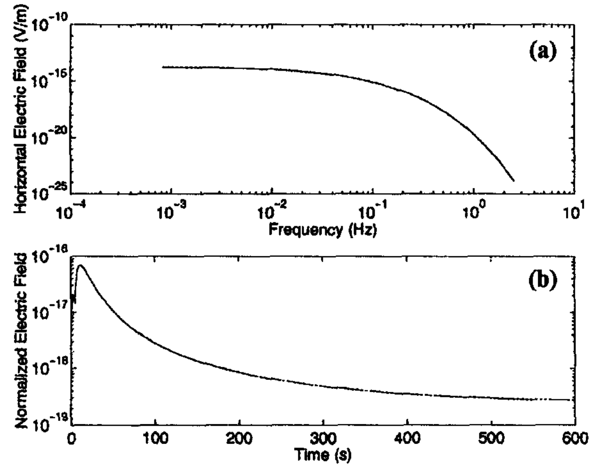

Fig. 2. a)The spectrum of the electric field of a dipole source, at a distance R=105km from the source. A Kronecker-δ source time function is assumed. The spectrum is normalized so that $E_H(100s)=1.12\times10^{-14}$V/m; see the text for an explanation. b) The waveform of the electric field with the spectrum of Fig. 2a.

from the source, only the very long periods may be detectable and the received waveform will, necessarily, exhibit a very slow variation with time.

Fig. 2 illustrates the received waveforms of transient fields from a single dipole source buried at d=10km in a 100Ωm halfspace and distance R=105km from the receiver. A Kronecker-δ source time function is assumed. Fig. 2a is the received spectrum and Fig. 2b the half width of the waveform, computed with direct numerical integration of the spectrum. The waveform exhibits a slow decay, decreasing by approximately one order of magnitude within less than 150s. This result is expected, since high frequencies cannot propagate far from the source, with consequence that the received spectrum becomes enriched in low frequencies and dies out slowly. This is an interesting result inasmuch as it shows how an instantaneous source may be stretched in time by many orders of magnitude, simply as an effect of its propagation over long distances through a finite conducting medium.

5 Propagation and waveforms of the field from multiple cracks (an excited cell).

In section 4, we have presented the properties of the field emitted by a single crack If, however, the cracks in a cell within the seismogenic volume are excited coherently, the resulting field will be the superposition of the fields from each individual crack. Consider a 1Am dipole source buried at d=5000m in a half space with ρ=100Ωm. If T=100s, then at R=100 km the total horizontal electric field is $E_h=1.12\times10^{-14}$ V/m. For an estimate of the actual current through an individual excited dipole crack, we have to rely on laboratory measurements. Warwick et al (1982) has found that $i_c\sim10^{-3}$ A. The lengths of microcracks and microfractures -clusters of dislocations - are of the order 10^{-4}-10^{-1} m (Molchanov and Hayakawa, 1995). We assume a mean length $l_c=10^{-3}$m. If a relatively rapid strain change occurs, then some elementary crustal volume will acquire new cracks. The maximum number of cracks containable in a unit volume is controlled by their size. Gershenzon et al (1989) provide the relationship $N_{MAX}=(3\cdot\tau\cdot\upsilon)^{-1/3}$ where N_{MAX} is the maximum crack density, τ is the time and υ is the velocity of crack opening. Assuming a constant velocity for crack opening $\upsilon=10^3$m/s, (comparable to seismic velocities), we find $\tau=10^{-6}$s and $N_{MAX}=3.7\times10^7$ m^{-3}. If N_{MAX} cracks are coherently excited in a volume $V\sim10^7$m^3 then at a distance of 100km the received horizontal electric field will be $E\sim N_{MAX}\cdot V\cdot i_c\cdot l_c\cdot E_h\sim4.1\times10^{-6}$ V/m = 4.1 mV/km. This is a correct order of magnitude, but also an upper limit. In order to obtain this amplitude (at T=100s), we require the maximum coherent excitation of a cell with dimensions 215×215×215 m.

A free parameter in this consideration, however, is our estimate of N. For instance, if $N\sim10^5$ and all other parameters are kept constant, the necessary dimensions of the excited cell (i.e. the effective volume V) will rise to 1km^3. We do not know what a realistic N may be and order of magnitude variations should be expected within the earthquake preparation zone. For instance, domains of increased fracturation (fault zone), or increased loading (asperities and barriers) may be expected to have higher N values. The distribution of strain is very complex within the earthquake preparation zone (Gershenzon et al, 1989), Thus, it should be expected that $N < N_{MAX}$ and changing with time; it may reach the maximum value only at the time of failure (Gershenzon et al, 1989). Accordingly, the effective volume required to produce a 4 mV/km amplitude at T=100s will be larger, or the signal weaker. It follows that for the MCD model to work, we require the collective or individual excitation of distributed cells with dimensions of the order of 1km^3 and strain rates of the order of 10^{-12}-10^{-11} s^{-1}.

Now, consider that a M5.5 earthquake has typical fault dimensions of 10×7 km and affects a crustal volume of 4.6×10^{11} m^3. It is not expected that the entire seismogenic zone will, or can be excited. Let us assume that conditions sufficient to trigger the generation of current are limited in a sub-volume in the immediate neighbourhood of the fault, say 10×7×1 km (7×10^{10} m^3). This is still is 10-10^2 times larger than the elementary cell sufficient to provide a 4 mV/km signal. In the case of a M6.5 event, these dimensions respectively become 30×12 km and the volume is 1.4×10^{13} m^3. Again, if conditions sufficient for excitation exist only in a sub-volume 30×12×5 km around the fault, this still is 10^3-10^6 time larger than that sufficient to give a 4mV/km variation. Therefore, it is conceivable that a sufficient number of cells above the effective volume threshold may appear in the seismogenic zone. However, if small cells are excited or small earthquakes concerned, signals may not be observed even at close ranges, because they're very weak to detect.

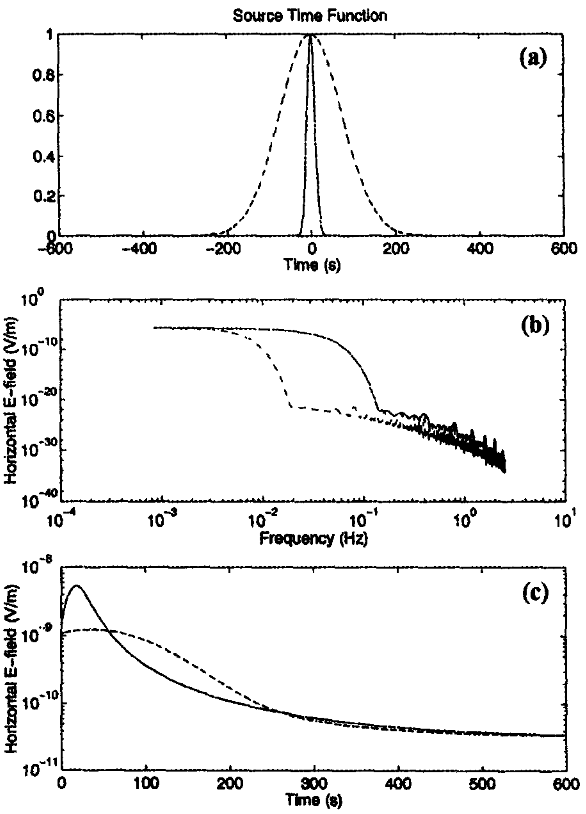

Fig. 3. a) The time function of a 'fast' (continuous line) and a 'slow' (dashed line) excited crustal cell. **b).** The spectra of the 'fast' (continuous line) and the 'slow' (dashed line) excited crustal cell, received at a distance R=105km. The spectra have been normalized so that E_H(T=1000s)=4×10^{-6} V/m. **c)** The waveform of the electric field emitted from the 'fast' (continuous line) and the 'slow' (dashed line) excited crustal cell, corresponding to the spectra of Fig. 3b.

One open question is that of the temporal behaviour of cell excitation and the duration of the source. We cannot provide some affirmative answer to this problem. Thus, and depending on conditions particular to the seismogenic zone in question, the entire zone may emit uniformly, or discretely in time (each cell on its own, when and if it is excited). The absolute lower limit for the duration of the source is τ (all the cracks in a cell emit simultaneously). In the general case, the duration will be a function of the pace at which strain builds up and propagates in the seismogenic zone. Consider however, that when strain increases, older cracks propagate and new cracks appear. Thus N increases and the effective excitation volume decreases. New, smaller cells may pass the excitation threshold and jump into action. This process will continue and may accelerate until the strain rate drops below the required threshold, upon which time it decelerates and stops. The duration of this process is unknown, but may, conceivably require several seconds or even minutes.

Let us, now, examine the waveforms of transient fields from an excited cell buried at d=10km and located at a distance R=105km from the receiver. The Earth has a uniform resistivity of 100Ωm. Fig. 3a shows the source time function. In order to study the effects of source spectrum on the propagation of the EM field, we consider two Dirac-δ types of source time functions : a faster with duration of 60s, which is richer in higher frequencies (continuous line) and a slower with duration 400s, richer in lower frequencies (dashed line). Fig. 3b shows the received spectrum, (the product of the spectrum of a unit amplitude Kronecker-δ type source, times the modulus of the spectrum of the time function, normalized to 4×10^{-6} V/m at T=1000s). Fig. 3c is the half-width of the corresponding time domain variation computed from the spectrum with direct numerical integration. The waveforms look like depressed and stretched replicas of the source time functions. When compared to each other, the faster source (continuous line) has a sharper onset, shorter duration and considerably faster decay than the slower source. These are effects of the source spectrum : the slower it is, the richer in low frequencies, which experience lower attenuation and dominate the received spectrum stretching the shape and the duration of the waveform. In order to further explore this point, we present Fig. 3d, illustrating the full 'received' waveform (dashed line) of a composite source function comprising a sharp onset and a slower decay (continuous line). Both the time function and the received waveform have been normalized to unity. Observe that the transient field has been stretched and smoothed as a consequence of high frequency attenuation, although some basic features of its original structure are still recognizable (its rise time is steeper than its decay time). The duration of the waveform is considerably longer than the duration of the source. In a concluding remark, we note that the simulated waveforms of Figures 2b and 3c,d exhibit several features reminiscent of some classes of bay-like variations, claimed to have been observed prior to earthquakes by several authors (e.g. Sobolev, 1975; Sobolev et al, 1986; Varotsos et al, 1993 and references therein.

6 Concluding remarks

In the foregoing, we have attempted to describe the pre-seismic transient current generation mechanism in terms of the Moving Charged Dislocation (MCD) model. We expand on the hypothesis that the source of a pre-seismic (or co-seismic) transient electric variation may be the generation of electric current in the earthquake focus due to the non-linear evolution of deformation during the preparation stage of an earthquake. Towards this effect, we have estimated the appropriate stress sensitivity coefficient, $E/(d\sigma/dt)$, necessary to initiate excitation of a rock volume. The resulting values are comparable with that calculated using the experimental data for different types of rocks. We provide a model which associates current density and the electric field with the rate of deformation ($\dot{\varepsilon}$) and we also

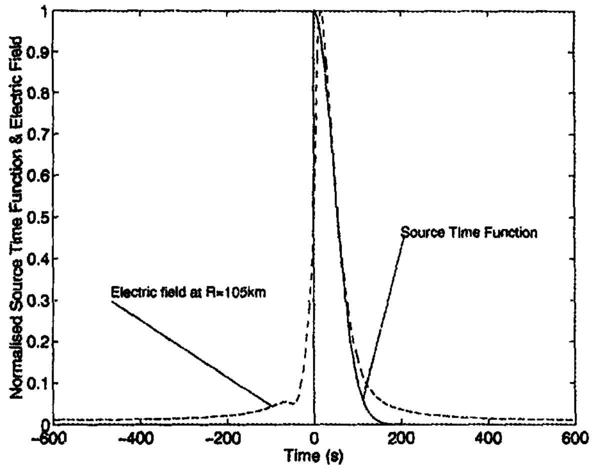

Fig. 3d. The time function (continuous line) and the 'received' waveform (dashed line), of a composite source with a sharp onset and a slower decay.

derive order of magnitude estimates of the strain rates necessary to initiate current generation. These estimates are high with respect to long-term geological strain rates. However, when 'normalized' by the fact that our calculations have been made on the basis of some drastic assumptions (homogeneous, linear, isotropic material), as well as laboratory data obtained under controlled conditions, the resulting estimates are not at all implausible. In fact, they are comparable to the observed strain rates of a large Japanese event. We believe that this result is significant in assessing the physical conditions appropriate for the generation of a pre-seismic signal and hence the validity of the MCD model.

We expand our exploration of the appropriate physical conditions of the earthquake preparation zone, by evaluating the size of the earthquake source volume capable of yielding a MCD signal observable at long distances from the source. We find that providing that the MCD model is capable of generating sufficient current in real Earth conditions, the observation of electric fields at intermediate - long distances is feasible for intermediate and large size events. Furthermore, we construct models of such waveforms which turn out to possess bay-like shapes, resembling some classes of signals observed by several independent investigators prior to earthquakes.

The model presented herein is by all means preliminary and certainly not proven. Although many of its components are demonstrable or physically arguable, there still exist several free parameters (the effective excitation volume at the earthquake zone to mention one), which require considerable effort before they can be tuned : the successful conclusion of such an undertaking is by no means certain. Nevertheless, we believe that we have presented enough evidence to demonstrate that the MCD process due to strain rate changes is geophysically plausible and a solid candidate generator of electrical earthquake precursors.

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