Tectonomagnetic modeling based on the piezomagnetism: a review

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Abstract
Development of tectonomagnetic modeling on the basis of the piezomagnetic effect is reviewed for the period since the early 1990's. First, the basic theory is briefly summarized, in which the representation theorem or the surface integral representation for the piezomagnetic potential and the Green's function method are presented. In the 1990's, several field observations in earthquakes and volcanoes were interpreted with the aid of analytic solutions based on the Green's function method. A general formula was developed for an inclined rectangular fault with strike-slip, dip-slip and tensile faulting. The surface integral method has been applied to 2D and 3D models, as well as to fault models in the inhomogeneously magnetized crust. When the magnetic field is measured within a bore hole, the effect of magnetic poles around the hole should be taken into account. As a result, tectonomagnetic signals are much enhanced in a bore hole compared with on the ground surface. Finally, piezomagnetic field changes associated with the Parkfield fault model are introduced and the new aspect of the model is discussed.

Key words *tectonomagnetic modeling – piezomagnetism – seismomagnetic effect of inclined faults – enhancement effect of bore hole – Parkfield fault model*

1. Introduction

The piezomagnetic effect of rocks was discovered in the 1950's (*e.g.*, Kalashnikov and Kapitza, 1952). The study of stress changes in the crust by magnetic observations is called *Tectonomagnetism*, which was coined by Nagata (1969). Tectonomagnetic modeling plays a key role in understanding the stress variation process within the crust by combining rock magnetic experiments and *in situ* field measurements. Some basic experimental results on stress-magnetization relationship were summarized by Nagata (1970).

The prototype of tectonomagnetic modeling was initiated by Frank D. Stacey in his famous paper «The Seismomagnetic Effect» (Stacey, 1964). Generally, mechanical models have complicated stress distribution, while the piezomagnetic law, or the stress-to-magnetization conversion, initially had a very complicated expression. Hence the piezomagnetic modeling has been done as a numerical computer simulation since the beginning. In particular, dislocation sources contain mathematical singularities: Even in the 1980's the reliability of numerical solutions was not clear.

A breakthrough was made not by the development of computers, but by reexamining Stacey's formulation. Bonafede and Sabadini (1980) and Sasai (1980) derived new formulas for the linear piezomagnetic effect independently of each other. They incorporated the elasticity theory of dislocations and derived an analytic solution for the piezomagnetic field produced by a vertical rectangular strike-slip fault. Their formulation is what is called the Green's function method.

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However, during the 1980's, it turned out that there were definite serious discrepancies among solutions by numerical volume integral and those by the Green's function method. The cause of the discrepancies was an enigma for about ten years. There were two typical examples: one is the Mogi model in volcanology which was discussed by Suzuki and Oshiman (1990) on the difference between Davis' (1976) and Sasai's (1979) solutions, while the other is the infinitely-long vertical strike-slip fault model on which Banks *et al.* (1991) examined Shamsi and Stacey's (1969) and Sasai's (1980) solutions. Finally, the cause of the discrepancy was identified: Some Green's functions by Sasai (1980) were not correct because of an inappropriate way of taking the limit of integrations around a point dislocation (Sasai, 1991b). The details of the controversy on this issue and how it was settled were described in Sasai's (1994b) review paper. I will present here the recent development of tectonomagnetic modeling studies since early 1990's.

2. Outline of the basic theory

The relation among stress components and magnetization under 3D stress state is given by Sasai (1980, 1991b):

$$\Delta \boldsymbol{J} = \frac{3}{2}\beta \boldsymbol{T}'\boldsymbol{J} \qquad (2.1)$$

$\boldsymbol{J}$ is the magnetization vector of the initial (stress free) state, while $\Delta\boldsymbol{J}$ its increment under the applied stress, and β is a material constant which is called stress sensitivity. $\boldsymbol{T}'$ is the stress deviation tensor, which is related to the stress tensor $\boldsymbol{T}$ as

$$\boldsymbol{T} = \sigma_0 + \boldsymbol{T}' \qquad (2.2)$$

where $\sigma_0 = (\tau_{xx} + \tau_{yy} + \tau_{zz})/3$ is the hydrostatic pressure.

With the aid of Hooke's law, we can express the stress-induced magnetization as a function of displacement

$$\Delta M_{kl} = \beta\mu\left[\frac{3}{2}\left(\frac{\partial u_k}{\partial x_l} + \frac{\partial u_l}{\partial x_k}\right) - \delta_{kl}\, div\, \boldsymbol{u}\right] J_k \qquad (2.3)$$

in which we denote $\Delta \boldsymbol{M}_k$ as the incremental magnetization produced by the k-th component of the initial magnetization $\boldsymbol{J}$ and ΔM_{kl} its l-th component.

The magnetic potential due to the piezomagnetization $\Delta \boldsymbol{M}_k$ is given by

$$W_k(\boldsymbol{r}) = \iiint_V \Delta \boldsymbol{M}_k(\boldsymbol{x}) \cdot \nabla\left(\frac{1}{\rho}\right) dV \qquad (2.4)$$

where $\boldsymbol{r}$ is the observation point, $\boldsymbol{x}$ an arbitrary point in V and $\rho = |\boldsymbol{r} - \boldsymbol{x}|$. Modifying eq. (2.4) with the aid of the Green's theorems, we can obtain the following expression (Sasai, 1983):

$$W_k(\boldsymbol{r}) = 4\pi C_k u_k(\boldsymbol{r})\Theta(\boldsymbol{r} \in V) + D_k \iiint_V \frac{F_k}{\rho} dV +$$

$$+ C_k \iint_S \left[\left\{-\frac{\partial u_k(\boldsymbol{r}')}{\partial n'} + \frac{2(\lambda+\mu)}{3\lambda+2\mu}\Delta \boldsymbol{m}^{(k)} \cdot \boldsymbol{n}'\right\}\frac{1}{\rho} + \right.$$

$$\left. + \{u_k(\boldsymbol{r}')\}\frac{\partial}{\partial n'}\left(\frac{1}{\rho}\right)\right] dS \qquad (2.5)$$

where

$$\Delta \boldsymbol{m}^{(k)} = \frac{3}{2}\left(\frac{\partial u_k}{\partial x_l} + \frac{\partial u_l}{\partial x_k}\right) - \delta_{kl}\, div\, \boldsymbol{u} \qquad (2.6)$$

and

$$\Theta(\boldsymbol{r} \in V) = \begin{cases} 1 & (\boldsymbol{r} \in V) \\ 0 & (\boldsymbol{r} \notin V) \end{cases}. \qquad (2.7)$$

Equation (2.5) is the representation theorem for the piezomagnetic potential produced by a homogeneous and isotropic elastomagnetic body. The first term on the righthand side of eq. (2.5) appears only when we observe the field within the magnetized body. The second term is the contribution from body force F_k. Usually it vanishes because we deal with tectonomagnetic models under no body force. Outside the body, the piezomagnetic potential is represented by the last term, *i.e.* a surface integral over the elastomagnetic body concerned. Obviously, this formula has a merit compared with the volume integral in eq. (2.4), because the integration is reduced from three to two dimensions.

As for the piezomagnetic potential produced by a dislocation surface within a semi-infinite elastic medium with a uniformly magnetized uppermost layer, the following formula was derived, *i.e.* Volterra's formula for piezomagnetic potential (Sasai, 1980, 1991b):

$$W^{m}(r)=\iint_{\Sigma}\Delta u_{k}(\xi)w_{kl}^{m}(\xi,r)v_{l}(\xi)d\Sigma(\xi) \qquad (2.8)$$

where $\Delta u_k(\boldsymbol{\xi})$ is the displacement discontinuity across the dislocation surface Σ and $\boldsymbol{\nu}$ the normal vector to Σ, respectively. And

$$w_{kl}^{m}(\xi,r)=C_{m}\iiint_{V}S_{kl}^{(m)}(\xi,x)\cdot\nabla\left(\frac{1}{\rho}\right)dV \qquad (2.9)$$

in which $S_{kl}^{(m)}(\boldsymbol{\xi},\boldsymbol{x})$ is the piezomagnetization associated with the m-th component of initial magnetization produced by a point dislocation specified by (kl). Hence w_{kl}^{m} is the piezomagnetic potential due to a point dislocation of the type (kl). We may call it the elementary piezomagnetic potential.

All these Green's functions w_{kl}^{m}'s were given explicitly by Sasai (1980). However, there were some serious discrepancies among the numerical models based on eq. (2.4) and analytical ones via eq. (2.8). Finally it was found that some of Sasai's (1980) Green's functions were not correct. More specifically, the integrand $S_{kl}^{(m)}(\boldsymbol{\xi},\boldsymbol{x})$ in eq. (2.9) has a singularity at the source position, and this improper integral is not uniformly convergent. We must choose an appropriate way of convergence in evaluating the integrals in eq. (2.9). Sasai (1991b) reexamined the integrals and obtained a revised version of w_{kl}^{m}'s. Analytical solutions based on the new Green's functions are consistent with those by numerical volume integrals (Sasai, 1994a).

3. Applicability of analytic solutions

Some of the tectonomagnetic observations have been successfully explained by analytical solutions of faults and pressure source models in a uniformly magnetized crust. Johnston *et al.* (1994) investigated coseismic magnetic changes associated with the Landers, California, earthquake of M_w 7.3 in 1992. About 1 nT changes in the total intensity were detected by careful data processing at two magnetometers within 20 to 30 km from the earthquake fault. Surface displacements could be modeled by a fault system with 11 segments, each of which was a vertical strike-slip fault. They showed that the total sum of contributions from each segment was coincident with the observation at both the magnetometer sites.

Some coseismic total intensity changes observed in Izu Peninsula, Central Japan, were also explained with the aid of the formula for a vertical strike-slip fault (Sasai and Ishikawa, 1997). In particular, a remarkable tectonomagnetic event took place in 1978, in which the total intensity gradually decreased by 7 nT since 3 months prior to a M 5 earthquake and it jumped up coseismically by 5 nT. The magnetometer was located just above a buried fault. The overall process of the magnetic changes was interpreted as due to aseismic slip of the fault at a shallow depth.

The total intensity change observed during the Matsushiro earthquake swarm (1965-1970) was the first reliable example of tectonomagnetic effects as revealed by proton precession magnetometers (Yamazaki and Rikitake, 1970). This observation was explained as a result of piezomagnetic effect (Sasai, 1994c). The multiple tension-crack model was applied to modeling the crustal deformation as well as magnetic changes in Matsushiro, which is a distribution of a number of small tensile cracks in an elastic half-space (Sasai, 1986).

Remarkable long-term magnetic variations associated with the inflation of resurgent domes in Long Valley caldera, California, since 1991 were explained by a hydrostatically pumped pressure source (the Mogi model) below the Curie point isotherm (Mueller *et al.*, 1991; Johnston, 1997; Sasai, 1991a). A more rapid volcanomagnetic effect (a few hours to days) observed at the time of the fissure eruption in Izu-Oshima volcano, Japan, in November 1986 can be explained by some vertical tensile faults (Sasai, in preparation).

A most general formula of the piezomagnetic potential due to an inclined fault with strike-slip, dip-slip and tensile fault motion was presented by Utsugi (1997) and Utsugi *et al.* (2000). Local absolute maxima in the magnetic fields appear above both the edges of the surface trace of the fault. The influence of fault inclination is as follows: the magnetic field becomes maximum when the fault inclination is 90° for strike-slip and tensile faults, as it does when the dip-slip fault is inclined by 45°. We can expect large coseismic changes for interplate earthquakes in subduction zones, but they would appear rather far off the coast. The computer code by Utsugi *et al.* (2000) enables us to easily estimate seismomagnetic or dyke-intrusion magnetic effects, and hence it would be of much help for us to design the optimal arrangement of magnetometers.

4. Development of the surface integral method

The numerical volume integral method based on eq. (2.4) has long been used for tectonomagnetic modeling since Stacey (1964). We have only to make the size of volume element small in order to attain enough accuracy of integration. In the case of the Mogi model, Suzuki and Oshiman's (1990) numerical solution coincides with Sasai's (1991a) analytic one within two to three figures. However, when the model includes any mechanical singularity, the convergency of integrals is not always guaranteed. The surface integral method was developed to analyse the behavior of integrals around a singular point (Sasai, 1991b). It is suitable to deal analytically with integration over a rather simple shape of envelopment around a singular point.

The surface integral method was first applied to numerical 2D modeling by Sakanaka *et al.* (1997). They successfully obtained the piezomagnetic field due to the Yukutake model, *i.e.* a horizontally embedded, infinitely long, pressurized cylinder in an elastic half-space (Yukutake and Tachinaka, 1967; Oshiman, 1990). They extended their model to the case that the ground surface had irregular topography of a volcano, in which they employed a numerical solution of the displacement field computed by the Boundary Element Method (BEM). Such a rugged surface produces local stress concentration, which results in a complicated distribution of piezomagnetic changes on the ground. Fluctuations of magnetic changes have almost the same amplitudes as the smooth variation in the case of the flat surface. This implies that the tectonomagnetic field could be locally enhanced owing to the irregular topography.

Sakanaka (1998) extended the numerical surface integration to 3D models. The curved boundary surfaces are approximated by the sum of small squares and triangles. He investigated in detail the piezomagnetic field produced by a finite cylindrical pressure source, whether vertical, inclined or horizontal, which was originally proposed by Walsh and Decker (1971) with application to the ground deformation in volcanoes. The next step would be to combine the surface integral method with BEM, in which we can utilize a variety of mechanical models. In the volume integral method, the use of FEM (Finite Element Method) had been already done by Zlotnicki and Cornet (1986) for 2D faults.

Utsugi (1999) applied the surface integral method to inhomogeneously magnetized crust with analytical approach. He obtained the piezomagnetic field produced by a uniformly magnetized cube under the stresses of an inclined fault; the fault motion can be strike-slip, dip-slip and tensile (Okada, 1992). He reduced the integrals over the cubic surface to 1D integrals which have integrands expressed by elliptic integrals. The 1D integral can be estimated accurately with the aid of the Double Exponential Formula (DEF). An important feature of Utsugi's (1999) result is that the solution becomes indeterminate when the cube involves the fault edge; in other words, the block must be subdivided so as to exclude the singular points. This is essentially the same difficulty as we met in the derivation of the elementary piezomagnetic potential (Sasai, 1991b, 1994b).

Utsugi (1999) estimated the possible seismomagnetic effect associated with an expected-to-occur earthquake in the central part of Japan, *i.e.* Tokai (or Suruga Bay) earthquake of M 8 or so. Only a few nT changes in the total intensity were expected in the case of the uniformly magnetized crust. However, the subducting slab (*i.e.* Izu Peninsula and its west- and downward extension) is strongly magnetized. With the aid of the cube formula, the model earthquake in the case of inhomogeneous magnetization was estimated to produce coseismic total intensity changes up to a few tens of nT.

5. Bore hole effect: enhancement of tectonomagnetic signals

Recently, the observation technique has been developed to set seismometer and/or strainmeter in a bore hole. One of the merits of the bore hole observations is that noises, in particular of artificial origin, are much reduced, and the other that it enables us to come closer to the sources of tectonic events. However, a physical quantity expressible by scalar potential such as gravity and magnetic field shows different behavior within its source material as compared with the case when we observe it outside the material. As for the piezomagnetic field, an additional term, *i.e.* the first term on the righthand side of eq. (2.5), appears. The origin of this term is as follows: When we observe within a magnetized body, we have to make a hole to put a magnetometer in it. Along the wall of the hole emerge magnetic poles, which produce the additional magnetic field. This effect can be regarded as the limiting case that we excavate a spherical hole and make its radius small.

However, the magnetic potential itself must be continuous across the material boundary. Across the boundary surface, it appears as if a gap in the potential value $4\pi C_k\{u_k(\boldsymbol{r})\}$ might occur owing to the vanishing of the first term in eq. (2.5) outside the body. This is completely compensated for by a jump in the potential across the double layer, *i.e.* the latter term in the surface integral of eq. (2.5), and hence the magnetic potential remains continuous. However, the normal component of the magnetic field varies discontinuously across the boundary, while the tangential ones are continuous. We have to take account of such a situation: The elementary piezomagnetic potential should be modified when the observation point lies under the ground (Sasai, in preparation).

Sasai (1994d) obtained coseismic magnetic fields to be observed under the ground associated with a vertical rectangular strike-slip fault. Figure 1a shows the computed piezomagnetic changes in the total intensity on the ground surface (2.5 m high), and fig. 1b underground

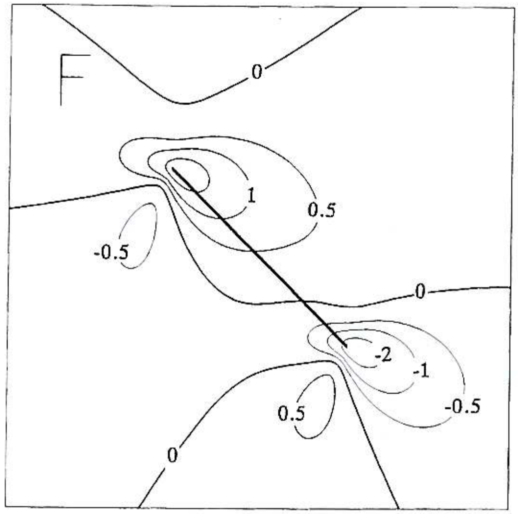

Fig. 1a. Computed piezomagnetic changes due to a vertical rectangular left-lateral strike-slip fault (the thick line). Contour lines show total intensity (nT) above the Earth's surface. After Sasai (1994d).

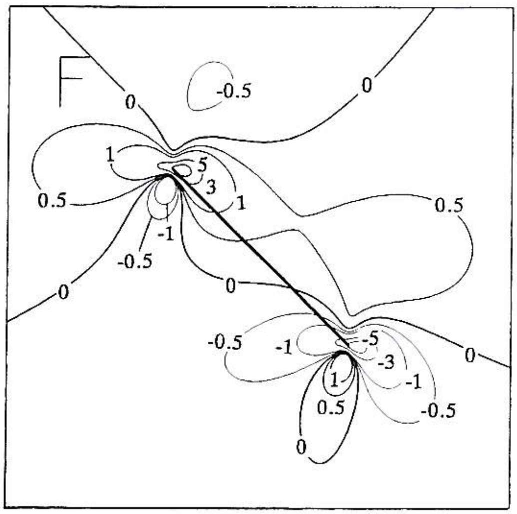

Fig. 1b. Computed piezomagnetic changes due to the same fault as fig. 1a. Contour lines show total intensity (nT) 100 m below the Earth's surface. After Sasai (1994d).

(at a depth of 100 m) due to an earthquake fault. The fault top lies at a depth of 500 m with its size 10 km long by 5 km wide and with left-lateral slip of 1 m. Generally, the total intensity changes under the ground are enhanced compared with the surface ones. The contribution from the vertical component is noticeable (Sasai, 1994d). The fields around both ends of the fault are large, where the effect of stress singularity also exists.

When the bore hole is filled with non-magnetic material (including air), magnetic poles emerge throughout the tube. As a result, the surface opening of the bore hole becomes sources and sinks of magnetic lines of force, which can extract some information of stress changes at depth (Johnston and Sasai, in preparation). When the casing pipe is made of high μ metal, the magnetic lines of force are absorbed and trapped within the pipe, but they spread out from the edge of the pipe: this might make things complicated. There still remain many problems with respect to the use of the bore hole. However, we can expect that it will make a breakthrough to improve the detection capability of weak tectonomagnetic signals by an order of magnitude.

6. Parkfield fault model: a new step

San Andreas fault is a transform fault running NW to SE in California, United States, which is a plate boundary between the North American and the Pacific plates. San Andreas fault system consists of some locked portions and always-creeping sections. Parkfield fault patch is located near around the boundary between the creeping section on the north and the locked patch of the 1857 Fort Tejon earthquake of M 8.3. Parkfield patch generates an earthquake of M 6 regularly with about a 30 year interval. The last earthquake occurred in 1966.

Stuart and Tullis (1995) made a computer simulation depicting how the fault patch slips slowly to arrive at the final break, in which they applied the fault friction law on the basis of the most recent model of earthquake generation process (Ruina, 1983; Tse and Rice, 1986). They subdivided the fault patch into a number of small rectangular fault segments. The surface displacements can be computed as the sum of contributions from these segments. In the same manner, the piezomagnetic field associated with this fault movement can be estimated as a sum of magnetic field from these small rectangular strike-slip faults. Stuart *et al.* (1995) conducted this computation, in which they used Sasai's (1991b) formula.

A merit of Stuart and Tullis' (1995) model is that the parameters of the fault constitutive law and their distribution on the fault surface were not given *a priori*, but were determined so as to fit the observed ground deformation around the fault. Although the «prediction» was not successful, *i.e.* the final rupture did not take place until the time they estimated, their model could be a prototype of earthquake prediction on the physical basis.

Stuart *et al.* (1995) presented the computed total intensity changes at some magnetometer sites near the fault during the entire earthquake cycle of 33 years or so. For example, at one site the coseismic step at the time of the 1966 earthquake is about 1 nT; it recovers and then builds up; and the same amount of coseismic decrease occurs at the next main shock. Another example is that they showed simple differences between the two points across the fault as we usually do in the field observations. Although the total intensity change is small, the reversal of general trend is expected to occur about five years prior to the main shock. Because of the drift-free nature of total intensity observations, such a trend reversal is easy to detect compared with strain measurements.

The Stacey model was the first step to the seismomagnetic calculations (Stacey, 1964). The use of the dislocation model was the second step (Shamsi and Stacey, 1969). Parkfield fault model would be the third step toward more realistic seismomagnetic effect. In Stuart *et al.*'s (1995) computation, only the measurement on the ground surface together with the uniform magnetization case is evaluated. If we take account of the measurement in bore hole as well as inhomogeneous magnetization, we can expect much more enhancement of seismomagnetic signals.

In conclusion, tectonomagnetic modeling theory can interpret actual field observations quantitatively in the 1990's after many efforts since the early 1960's. Tectonomagnetic observation is important, because it can provide us with information on stress changes within the crust independently of mechanical means. One of our major subjects is now to improve the detection capability of tectonomagnetic signals.

Acknowledgements

Discussions with Mitsuru Utsugi and Shin'ya Sakanaka were very helpful to complete this review. I am greatly indebted to every suggestion by and discussions with Naoto Oshiman, Yoshikazu Tanaka, Malcolm Johnston, Jacques Zlotnicki, Hisashi Utada and Yasunori Nishida.

REFERENCES

BANKS, P.O., W.D. STUART and S.-W. LIU (1991): Piezomagnetic fields of screw dislocation fault models, *J. Geophys. Res.*, **96**, 21575-21582.

BONAFEDE, M. and R. SABADINI (1980): A theoretical approach to the seismomagnetic effect, *Boll. Geofis. Teor. Appl.*, **22**, 105-116.

DAVIS, P.M. (1976): The computed piezomagnetic anomaly field for Kilauea Volcano, Hawaii, *J. Geomagn. Geoelectr.*, **28**, 113-122.

JOHNSTON, M.J.S. (1997): Review of electric and magnetic fields accompanying seismic and volcanic activity, *Surv. Geophys.*, **18**, 441-475.

JOHNSTON, M.J.S., R.J. MUELLER and Y. SASAI (1994): Magnetic field observations in the near-field of the 28 June 1992 M_w 7.3 Landers, California, earthquake, *Bull. Seismol. Soc. Am.*, **84**, 792-798.

KALASHNIKOV, A.G. and S.P. KAPITSA (1952): Magnetic susceptibility of rocks under mechanical stresses, *Dokl. Akad. Nauk. USSR*, **86**, 521-523.

MUELLER, R., M.J.S. JOHNSTON and J. LANGBEIN (1991): Possible tectonomagnetic effects observed from mid-1989 to mid-1990, in Long Valley Caldera, California, *Geophys. Res. Lett.*, **18**, 601-604.

NAGATA, T. (1969): Tectonomagnetism, *IAGA Bull.*, **27**, 12-43.

NAGATA, T. (1970): Basic magnetic properties of rocks under the effect of mechanical stresses, *Tectonophysics*, **9**, 167-195.

OKADA, Y. (1992): Internal deformation due to shear and tensile faults in a half-space, *Bull. Seismol. Soc. Am.*, **82**, 1018-1040.

OSHIMAN, N. (1990): Enhancement of tectonomagnetic change due to non-uniform magnetization in the Earth's crust – two dimensional case studies, *J. Geomagn. Geoelectr.*, **42**, 607-619.

RUINA, A.L. (1983): Slip instability and state variable friction laws, *J. Geophys. Res.*, **88**, 10359-10370.

SAKANAKA, S. (1998): Tectonomagnetic models evaluated by numerical surface integrals, *Ph.D. Thesis*, Kyoto University, pp. 66.

SAKANAKA, S., N. OSHIMAN and N. SUMITOMO (1997): A hybrid calculation method of tectonomagnetic effect using BEM and the surface integral representation of the piezomagnetic potential – Two dimensional case study, *J. Geomagn. Geoelectr.*, **49**, 101-118.

SASAI, Y. (1979): The piezomagnetic field associated with the Mogi model, *Bull. Earthquake Res. Inst., Univ. Tokyo*, **54**, 1-29.

SASAI, Y. (1980): Application of the elasticity theory of dislocations to tectonomagnetic modelling, *Bull. Earthquake Res. Inst., Univ. Tokyo*, **55**, 387-447.

SASAI, Y. (1983): A surface integral representation of the tectonomagnetic field based on the linear piezomagnetic effect, *Bull. Earthquake Res. Inst., Univ. Tokyo*, **58**, 763-785.

SASAI, Y. (1986): Multiple tension-crack model for dilatancy: surface displacement, gravity and magnetic changes, *Bull. Earthquake Res. Inst., Univ. Tokyo*, **61**, 429-473.

SASAI, Y. (1991a): Piezomagnetic field associated with the Mogi model revisited: analytic solution for finite spherical source, *J. Geomagn. Geoelectr.*, **43**, 21-64.

SASAI, Y. (1991b): Tectonomagnetic modeling on the basis of the linear piezomagnetic effect, *Bull. Earthquake Res. Inst., Univ. Tokyo*, **66**, 585-722.

SASAI, Y. (1994a): Resolution of contradiction between seismomagnetic models, *J. Geomagn. Geoelectr.*, **46**, 329-340.

SASAI, Y. (1994b): Piezomagnetic fields produced by dislocation sources, *Surv. Geophys.*, **15**, 363-382.

SASAI, Y. (1994c): A generating mechanism of Matsushiro, Central Japan, earthquake swarm: spontaneous hydraulic fracturing, in *Proceedings Conductivity Anomaly Symposium (1994)*, 181-195 (in Japanese).

SASAI, Y. (1994d): Enhancement of piezomagnetic signals within a bore-hole, in *Electromagnetic Phenomena Related to Earthquake Prediction*, edited by M. HAYAKAWA and Y. FUJINAWA, 51-54.

SASAI, Y. and Y. ISHIKAWA (1997): Seismomagnetic models for earthquakes in the eastern part of Izu Peninsula, Central Japan, *Ann. Geofis.*, **40** (2), 463-478.

SHAMSI, S. and F.D. STACEY (1969): Dislocation models and seismomagnetic calculations for California 1906 and Alaska 1964 earthquakes, *Bull. Seismol. Soc. Am.*, **59**, 1435-1448.

STACEY, F.D. (1964): The seismomagnetic effect, *Pageoph*, **58**, 5-22.

STUART, W.D. and T.E. TULLIS (1995): Fault model for preseismic deformation near Parkfield, California, *J. Geophys. Res.*, **100**, 24079-24099.

STUART, W.D., P.O. BANKS, Y. SASAI and S-W. LIU (1995): Piezomagnetic field for Parkfield fault model, *J. Geophys. Res.*, **100**, 24101-24110.

SUZUKI, Y. and N. OSHIMAN (1990): A paradox in volcanomagnetism: disagreement between analytical and numerical estimates of geomagnetic changes due to an underground pressure nucleus, *J. Geomagn. Geoelectr.*, **42**, 1291-1308.

TSE, S.T. and J.R. RICE (1986): Crustal earthquake instability in relation to the depth variation of frictional slip properties, *J. Geophys. Res.*, **91**, 9452-9472.

UTSUGI, M. (1997): A computer program for the calculation of geomagnetic field changes due to an inclined rectangular fault in a half space, *J. Fac. Sci., Hokkaido Univ.*, Ser. VII, **10**, 295-310.

UTSUGI, M. (1999): A theoretical study on seismomagnetic effect considering the inhomogeneously magnetized Earth's crust, *Ph.D. Thesis*, Hokkaido Univ., pp. 128.

UTSUGI, M., Y. NISHIDA and Y. SASAI (2000): Piezomagnetic potentials due to an inclined rectangular fault in a semi-infinite medium, *Geophys. J. Int.*, **140**, 479-492.

WALSH, J.B. and R.W. DECKER (1971): Surface deformation associated with volcanism, *J. Geophys. Res.*, **76**, 3291-3302.

YAMAZAKI, Y. and T. RIKITAKE (1970): Local anomalous changes in the geomagnetic field at Matsushiro, *Bull. Earthquake Res. Inst., Univ. Tokyo*, **48**, 637-643.

YUKUTAKE, T. and H. TACHINAKA (1967): Geomagnetic variation associated with stress change within a semi-infinite elastic earth caused by a cylindrical force source, *Bull. Earthquake Res. Inst., Univ. Tokyo*, **45**, 785-798.

ZLOTNICKI, J. and F.H. CORNET (1986): A numerical model of earthquake-induced piezomagnetic anomalies, *J. Geophys. Res.*, **91**, 709-718.