MAGNETIC FIELD ANOMALIES ASSOCIATED WITH GEODYNAMIC PHENOMENA

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Abstract. Significant variations of the Earth's magnetic field seem to be associated with geodynamic phenomena, particularly volcanic eruptions and earthquakes. The variations have small amplitudes, generally only a few nT. These small signals are embedded into larger transient variations due to other natural or artificial causes. An original method of recovering interesting signals having amplitudes down to 1 nT, is to describe the whole variation analytically so that each component may be isolated. Several studies have shown that significant signals exist for tectonically active regions: 10–30 nT for large-scale arrays over periods of years; 5–15 nT for volcanic phenomena; 1–5 nT for earthquakes.

Only a few test sites for volcanoes (New Zealand, Hawaii, Antilles) and earthquakes (essentially the San Andreas fault system) have been monitored so far.

The cause of the type of magnetic variations studied here, which are associated with volcanism and rock failures, has most often been attributed to piezomagnetism. Some authors have suggested that electrokinetic effects associated with porosity-changes may also play a role.

The isolated signals may, in the future, be used for hazard forecasting when the time-constants associated with the effects are known more precisely.

1. Introduction

The term 'tectonomagnetic effect' is usually applied to seismomagnetic effects, even though earthquakes are only one of the tectonic phenomena. This term was proposed by Nagata (1969). Records of magnetic effects supposed to be due to earthquakes data back to the late 1700 s. In 1890, the British seismologist, John Milne, was one of the first to review these reports. However, in 1914, H. F. Reid could show that these earlier reports were incorrect; magnetic effects were simply due to the magnetometer jiggling (up and down) as the seismic waves passed by. The results from 1890 to 1967 have been evaluated by Rikitake (1976) who found most of them unconvincing; in fact, the better the measuring technique, the smaller was the magnitude of the magnetic field-change associated with earthquakes. This is illustrated in Figure 1.

Thus, taking account of these results, it is necessary to study first the noise and its origin, then to examine the most recent available results. Later, we will see the geodynamic significance of these results and their implications. Finally, we will briefly describe the principal mechanisms involved in order to explain the observed values.

2. Noise Elimination

Since about 1960, several investigations have considered the difficulties that exist in isolating a tectonomagnetic signal as low as 1 nT from a noisy signal often showing 10

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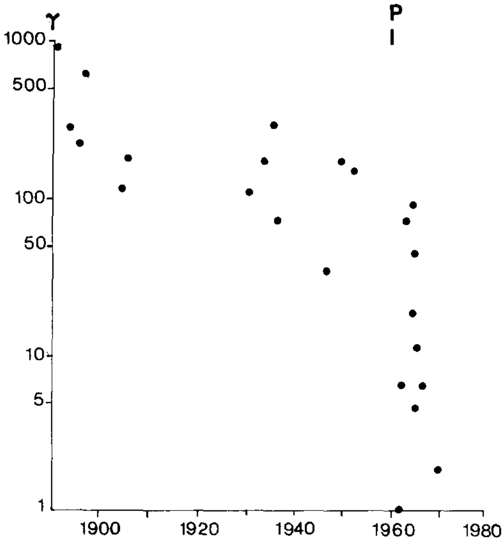


Fig. 1. Variation with year of geomagnetic changes associated with earthquakes. *P* represents the epoch when the proton magnetometer was introduced to geomagnetic field work. (After Rikitake, 1976).

times greater variations. Two different approaches have been followed: the first proposed by Rikitake (1966) is a statistical one, the second proposed by Johnston *et al.* (1976) and Pozzi *et al.* (1979) is an analytical one.

Rikitake (1966) shows that reduction of the observations is not sufficient to eliminate external transient variations completely. He compares the total intensity values of the geomagnetic field (F) observed at Matsushiro (MAT) where a violent earthquake swarm occurred in 1965–1967 with those recorded at Kanozan (KAN) about 200 km from Matsushiro. However the values of differences scattered with a standard deviation of about 3 nT. A weighted-difference technique is proposed for noise elimination, of the form $F_{\mathrm{MAT}} - CF_{\mathrm{KAN}}$ (where C is an empirically-determined constant. In this particular case, this does not reduce statistical deviations to less than 2 nT.

The reasons may be found in two aspects of the observations. First, total field differences between MAT and KAN stations have a large dip-component because of geographical location and, second, day and night data were not analyzed independently. Scattering of dip differences is known to be larger for daytime data. The observations of Mori and Yoshino (1970) on relations between daytime and night-time relations apparently agree with these comments.

The method of analysis of Rikitake has been used by Ispir and Uyar (1971) and Isikara *et al.* (1979) for studies of tectonomagnetism in Turkey (see Section 3.2).

The second way to reduce the noise experimentally is to use differential measurements with short base lines (no more than 10 to 20 km). The standard deviation of the differences then becomes less than 1 nT. In order to reduce the background noise further, it is necessary to understand the sources of this noise. An attempt in this direction has been made by Johnston *et al.* (1976), who isolate two main noise sources. First, a small-amplitude, high-frequency incoherent noise with about 0.5 nT standard deviation, present in all differences at all times. This noise can easily be reduced by simple averaging techniques. Superimposed on this, a larger amplitude, generally coherent low-frequency sporadic noise, clearly related to ionospheric and magnetospheric disturbances. The amplitudes of these disturbances for each difference appears to be a function of (1) station separation, (2) bed rock susceptibility contrast, (3) electrical conductivity structure.

For these authors, the most important effects are 1 and 2. They conclude that the detection capability at the 0.5 nT level or slightly better for intermediate to long-period tectonomagnetic effects does appear possible at most sites in the array studied in California. Shorter-period effects (less than a day) will be difficult to recover unless they exceed three or four nT.

A more recent paper in which this topic is mentioned is that of Pozzi *et al.* (1979). The authors separate the geomagnetic field into four parts: (1) the main one which varies slowly in time and space and is subject to secular variation which can be taken as uniform at the scale of 30 to 50 km, (2) an anomalous field assumed to be constant in time but rapidly varying in space, (3) a transient field which varies both in time and space, (4) the geodynamic magnetic effect which is still unproved.

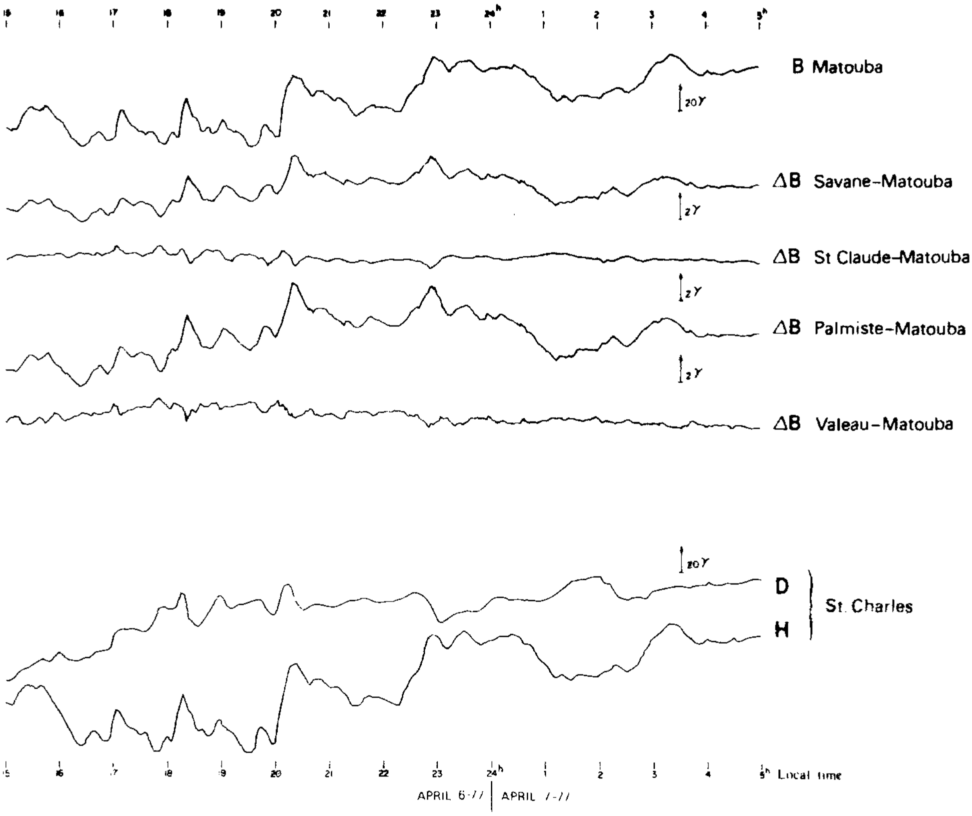


Fig. 2. An example of difference between field changes observed in stations over the Soufrière Massif. The location of the stations is presented in Figure 12. (After Pozzi *et al.*, 1979).

Examining the differences, it appears that errors can be classified into categories with known properties.

First, the projection error due to differences in dip of the Earth's magnetic field, F, at different stations. This kind of error can be avoided by choosing stations where values of elements of the Earth's magnetic field are not too different. The projection error varies like normalized differences in values of F. This term can be important when stations are far from one another and may also partly explain the constant coefficient between two differences as observed, for example, by Rikitake (1966).

Second, the conductivity anomaly error linked to differences of external origin for which internal electrical currents can be locally concentrated in conductive structures as has been shown in different areas of the world (as will be shown later). As in the preceding case, corrections are generally possible by studying the anomalous transient field at every point of the survey and determining the transfer coefficient from one station to another, as shown in Figures 2 and 3. Note that this kind of correction is applied only when transient variations are important.

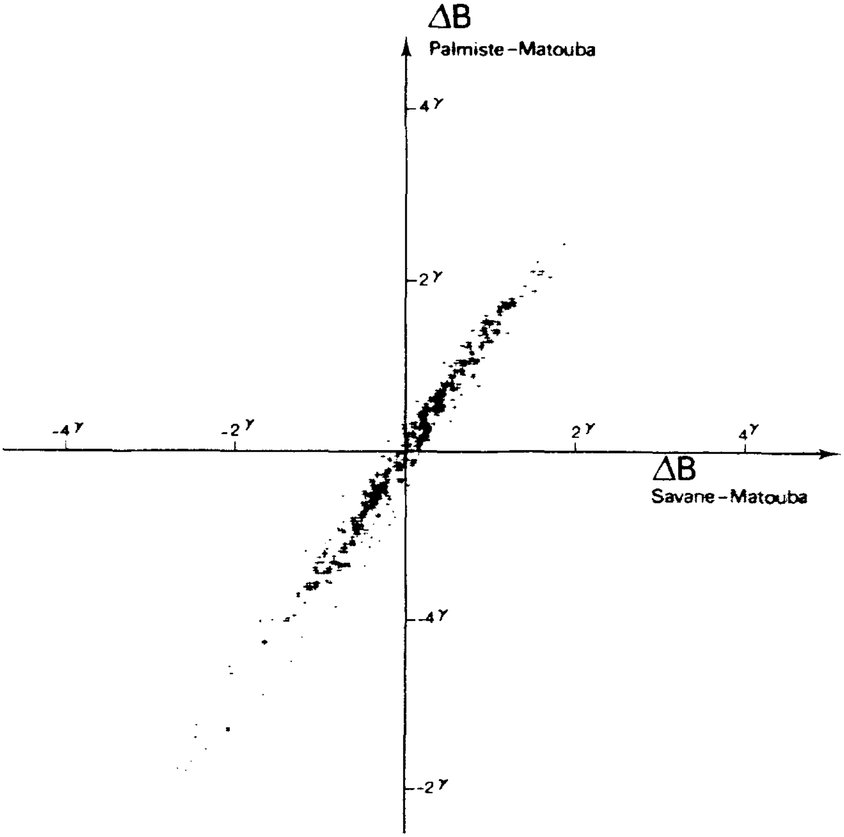


Fig. 3. Variation of a difference against another difference over the Soufrière Massif (Guadeloupe). (After Pozzi *et al.*, 1979).

Third, if the distance between two stations is sufficiently large, it is also necessary to correct for changes in the secular variation; however, this kind of correction is often unnecessary, taking into account the limited duration of surveys.

We have seen how many difficulties exist in isolating magnetic effects related to geodynamic phenomena, either earthquakes, volcanic eruptions, or strain-stress modifications in active geodynamic regions.

We will now summarize the most important results which have been described in this domain. However, we should like to mention that some important results are unpublished either because they were negative (e.g. experiments between two phases of the Soufrière of Guadeloupe during the years 1964–1965), or because measurements are made mostly by applied research institutes and are published only in internal reports. The results that we present are essentially only published results and thus most are positive.

3. Experimental Results

It seems possible to obtain a significant signal linked to geodynamical processes, either volcanic or seismic or more generally tectonic as we have seen in Section 2. We shall now examine the results as functions of time and space; it is possible to form three classes:

First: variations with long time-constants which mostly depend on relations existing between geodynamics and modifications of the Earth's magnetic field. The explanation of these phenomena necessarily had to be included in global models for geodynamics.

Second: variations with periods of days or months corresponding mostly to modifications of stress in the studied areas.

Third: there exist variations with periods shorter than days that do not seem to correspond to the phenomena studied here. Nevertheless, these variations seem directly connected with tectonics and more general geodynamics: for example, the conductivity anomaly of the Pyrénées in France (Galdeano *et al.*, 1979) or the conductivity anomaly in Central Europe (Singh, 1978) or along the Western part of the U.S.A. (Schmucker, 1970).

In the present paper, we shall only look at the results which are directly related to active geodynamic phenomena such as volcanoes or earthquakes.

3.1. LONG TIME-CONSTANT EFFECTS

It appears that there is a correlation between regions under strain (like foldings) and slow variations of the Earth's magnetic field. Some observations of this kind of phenomenon have been made in Japan, U.S.S.R. and GDR.

From about 25 yrs of absolute measurements of the magnetic components in G.D.R., Mundt (1978) has pointed out the existence of anomalies of the secular changes. These results indicate anomalies of 3–4 nT yr^{-1}; most of them have been noticed over regions where recent vertical and horizontal movements have been observed. Moreover, they seem to be associated with positive anomalies of heat flow. This aspect is shown in Figure 4 where the relation between anomalies of heat flow and secular variation is presented.

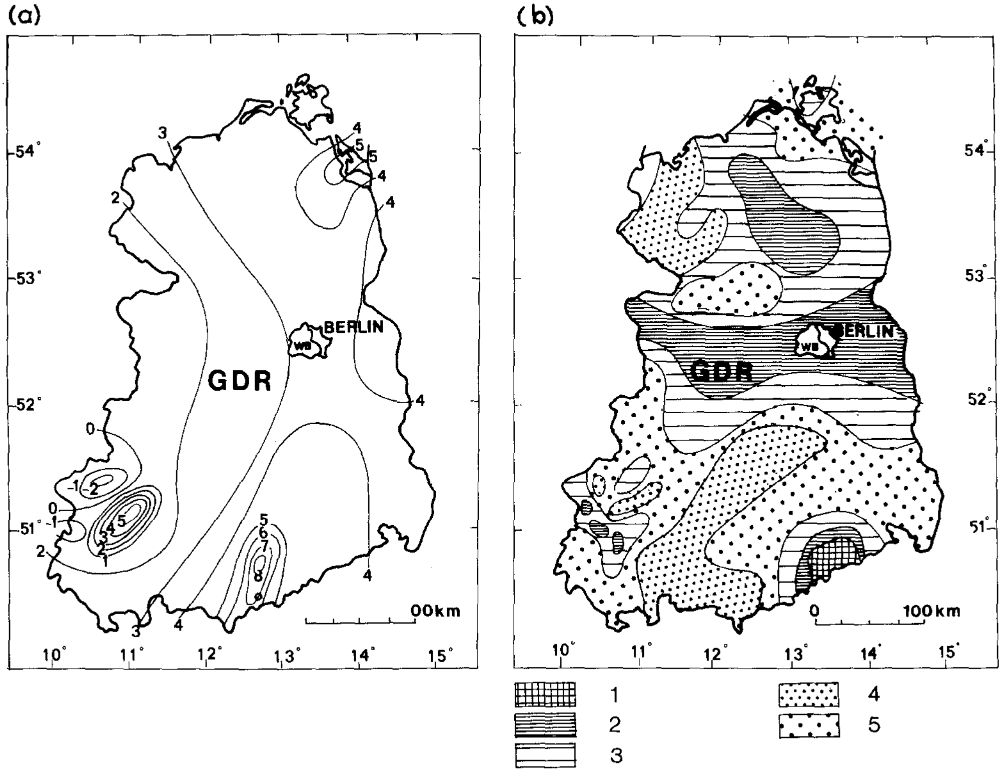


Fig. 4a. Secular variation anomalies $\delta\Delta F$ in nT year. The trend is a one degree polynomial (after Mundt, 1978). (b) Regional pattern of the heat flow (HF) in μ cal cm^{-2} s; (1) >2,0; (2) 1,8<HF<2,0; (3) 1,6 H HF H 1,8; (4) 1,4 HF 1,6; (5) HF 1,4. (after Hurtig in Mundt, 1978).

In a first stage of interpretation, it seems possible to relate modifications of the Curie isotherm with the observed effects linked to tectonics. Taking into account local movements, it is also probable that magneto-mechanical phenomena play some role.

In Japan, local anomalies of the secular variation have also been observed (Figure 5). Tazima *et al.* (1976) have shown by means of isoporic charts that anomalies of 2 nT/yr^{-1} or more seem to be correlated with tectonic activity such as earthquakes and volcanism. There exists a remarkable anomaly in the southeastern part and western coast area of Hokkaïdo. The former anomaly might have something to do with a large earthquake ($M = 7.4$) that occured off the Pacific coast in June 1973. In this study, it also appears that the observed anomalies are explainable with thermal modifications (volcanism) and with mechanico-magnetic modifications (earthquakes) linked to changes in the regional stress.

Modifications of the secular variation observed by Nagata (1976) in the neighbouring regions of the western coast of Japan (Figure 6) may be related to the Niigata earthquake of 1964. Other observations of this kind have been made in U.S.S.R. areas, particularly by Shapiro *et al.* (1978). These authors have pointed out a relation between secular

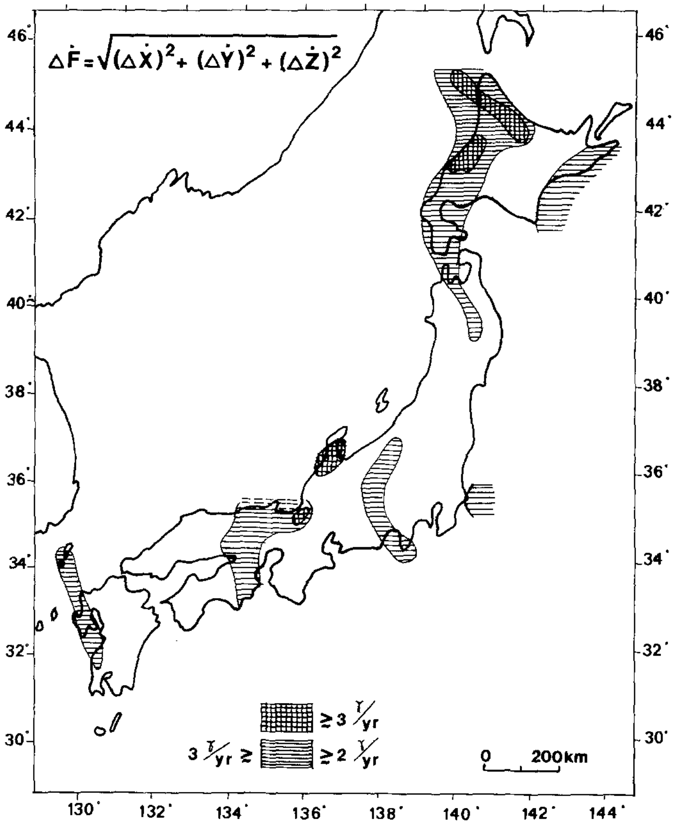


Fig. 5. Distribution of the resultant magnitude of local magnetic secular change anomalies (after Tazima *et al.*, 1976).

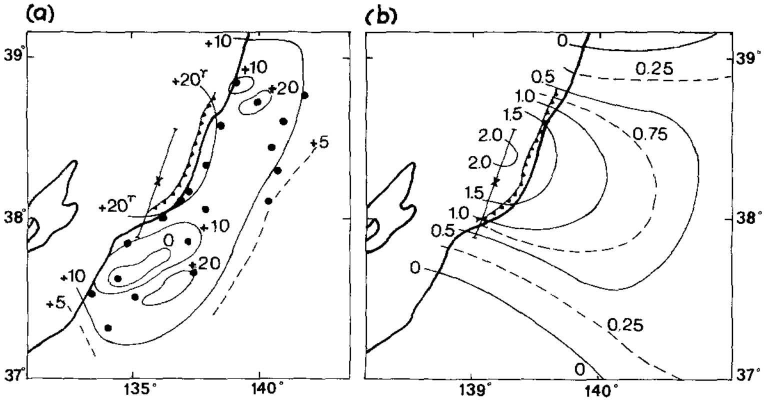


Fig. 6a. Locally anomalous secular variation in the geomagnetic vertical field during the prequake (1954–1961) observed in the neighbourhood of the focus of Niigata earthquake. (b) Locally anomalous secular variation in the vertical field as expected from the tectonomagnetism theory (in units of $B\, G_0\, J_0$) after Nagata, 1976).

variation and aseismic geodynamic regions in the Urals. The aim of their study was to try to determine secular variation anomalies from over 3500 stations and to study their relation to geodynamic processes. Such processes in tectonically active zones of the lithosphere may give rise to a local distortion of the normal pattern of geomagnetic field variations observed at the Earth's surface as has been shown by Shapiro (1976).

Long time-constant variations are poorly described in the literature because they need long temporal series with appropriate sampling rates. In our opinion, this kind of phenomenon must exist, but its amplitude is probably lower that those described. This magneto-geodynamic effect can be explained by three types of mechanisms: (1) thermal, (2) piezo-magnetic, (3) electrokinetic.

3.2. TECTONOMAGNETIC OR SEISMOMAGNETIC EFFECTS

So far, only a few areas of the world have been investigated, essentially Turkey, U.S.S.R., U.S.A. Some observations seem to have been made in China, but the results are still poor or unpublished.

The San Andreas fault system (California, U.S.A.) seems to be the best monitored region in the world.

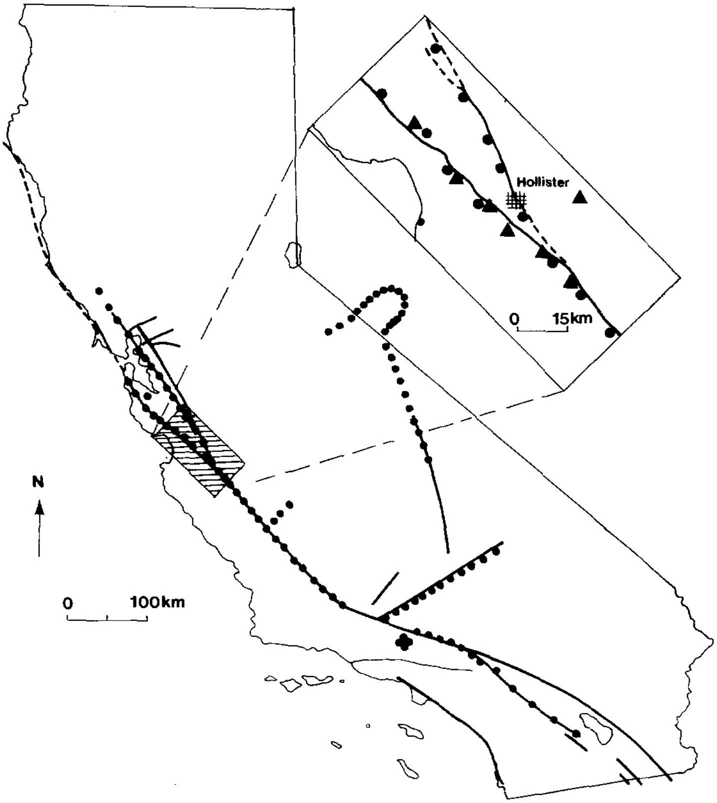


Fig. 7. Location of stations along the San Andreas fault, (after Johnston *et al.*, 1976).

California is a very good laboratory for seismic surveillance. Many experiments are done for deformations, stress and seismicity all along the San Andreas fault (see, Figure 7). For measurements of the magnetic field, magnetometers are operated a few times a year by Johnston and his team at more than 150 sites along 1200 km, by occupying two sites at a time, about 75 synchronized readings are taken over a 10 min period when the geomagnetic field is relatively undisturbed.

Few significant results have been described. The data from 12 sites, about 12 km apart, on the Garlock fault have revealed the existence of a magnetic effect related to a series of earthquakes of magnitudes 3.8 and 4.3 that occured near sites 9 and 10 (Figure 8). In this figure, it is shown that the magnetic field changes at various intervals of time as a function of site. It is evident that for sites 9 and 10, the signal is quite significant. Other evidence for a tectonomagnetic effect has been shown in relation to an earthquake of magnitude 5.2 in Central California near Hollister. Taking into account the noise level (0.25 nT), the signal of more than 0.5 nT between stations 4 and 5 is significant. The results are presented in Figure 9.

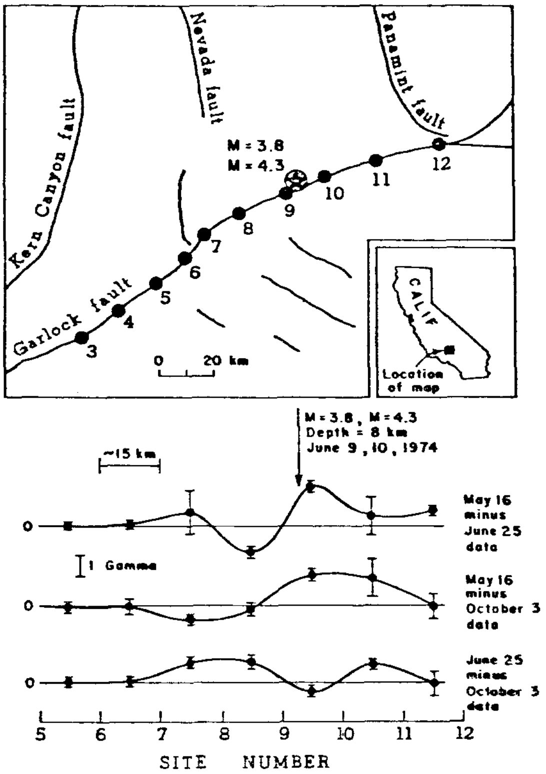


Fig. 8. Variation of the magnetic field on the Garlock fault for various intervals from May to October 1974. Earthquakes of magnitude 3,8 and 4,3 occured on June 9 and 10, (after Johnston *et al.*, 1976).

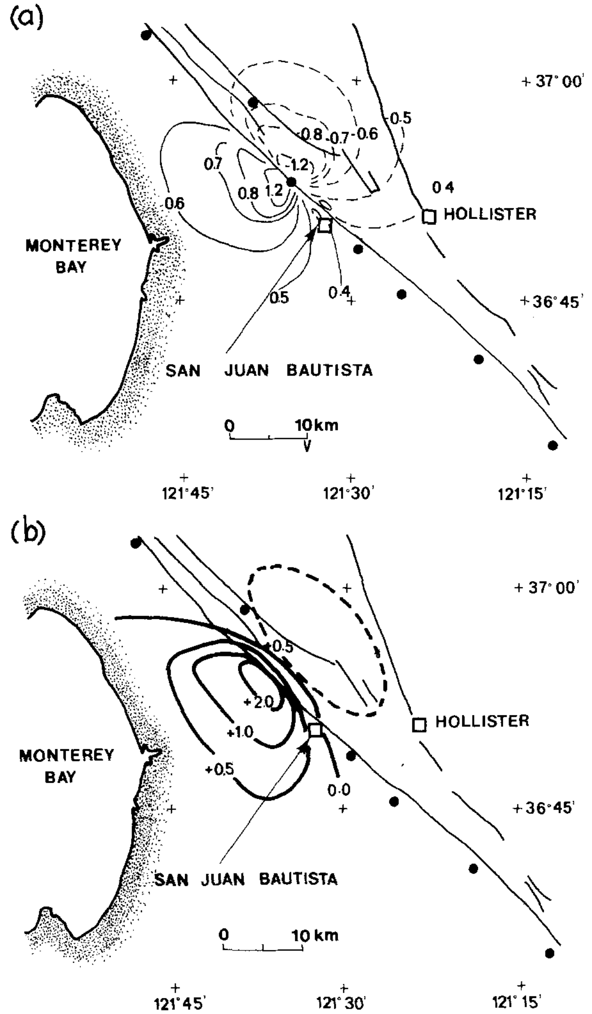
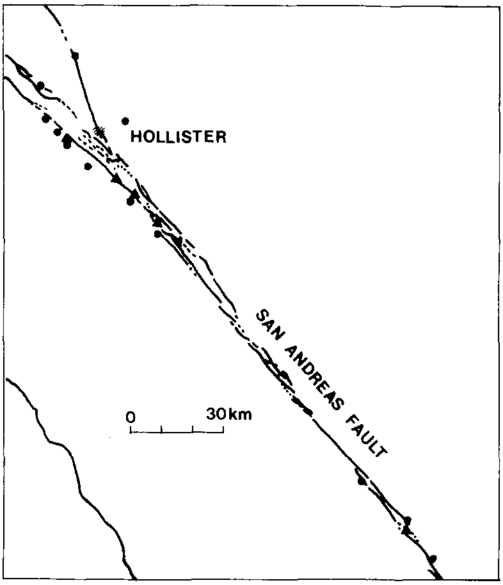


Fig. 9a. Magnetic changes related to Hollister region earthquakes. (b) Observed and calculated effect near San Juan Bautista Hollister (after Johnston, 1978).

For most of the authors, the mechanism most often invoked is a modification of magnetization (by pressure) but it does not seem that any definite relationship between magnetic changes and earthquakes has yet been established.

The San Andreas fault system has been the object of other similar studies. Johnston (1978) points out a relationship between stress changes along a slip zone and the local magnetic field variation. Slip along a limited section of the zone N of the San Andreas fault apparently produced an anomaly of 2 nT during a 5 yr period. The effect was attributed by him to anomalous secular variation or to tectonomagnetic induction.

Breiner and Kovach (1968), Smith *et al.* (1978), and Johnston (1980) have looked for relations between magnetic effects and fault creep observations in other areas. It seems that although some patterns may appear (see Figure 10), there is no significant correlation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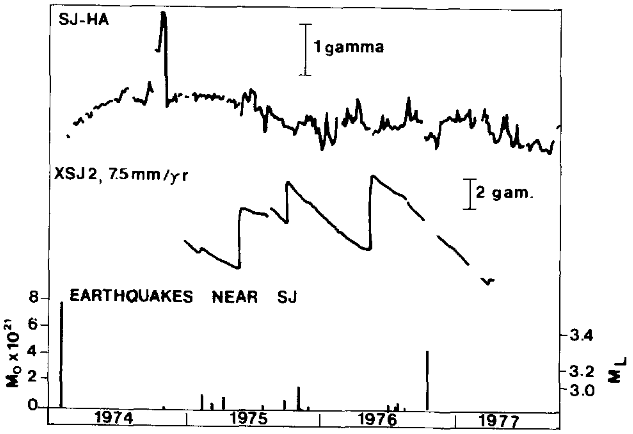
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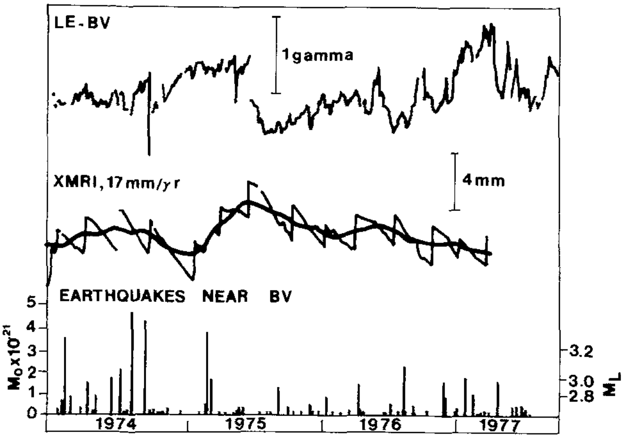
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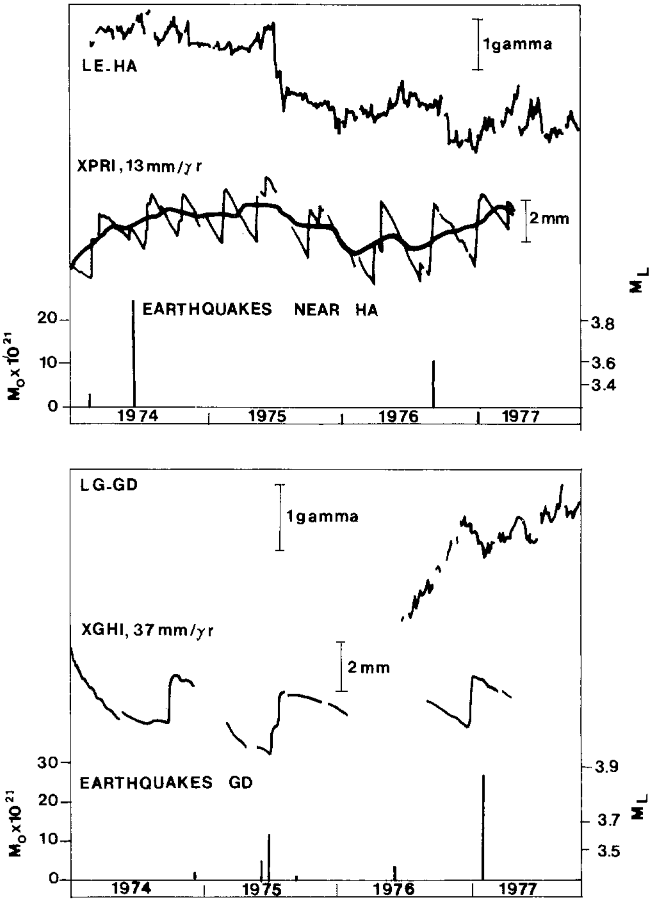

Fig. 10. Relation between creep effects and magnetic variations (after Smith *et al.*, 1978).

between groups of deep events and magnetic changes. It is also likely that possible correlations depend on other parameters in this complex domain. The need for long temporal series of data for the study of the relationships is quite apparent.

We think that the latter experiments are more continuous and more precise for the seismomagnetic effect. The effects pointed out in this region are very weak (less than 5 nT) and appear more and more weak with the improvement of observational techniques.

It is also necessary to mention the study of Madden (1980) in this region which is discussed by D. Beamish in the same issue of this journal.

Ispir and Uyar (1971), using the weighted technique to isolate a seismomagnetic signal, have pointed out that there exist some variations in the differences between two stations situated about 100 km apart. They attribute this effect to earthquakes and they think that there exists a relation between the location of the earthquake and the observed effect.

Isikara *et al.* (1979) working in Balkan countries and using a weighted difference technique and taking the ratio of components, present some results obtained after the

Bucarest earthquake.

In middle Asia, many investigations have been made by Shapiro *et al.* (1978) and Shapiro and Abdullabekov (1978). They describe some effects up to 25 nT related to an earthquake in the Tashkent region. It also appears that some effects exist in relation with earthquakes in the region of Kysyl Kum. Two interesting new experiments have been done, one near the Charvak reservoir where they observed a magnetic effect with a time constant of about three months. This effect, which can reach up to 10 nT, is attributed to the pressure effect and to the electrokinetic effect. Another interesting effect of 10 nT has been observed during the Gazly earthquake of 17 May 1976. It has been measured 29 h before the principal earthquake. It is also interesting to notice that no effect was observed during 'aftershocks'. For this last event, it is interesting to notice that there exists a conductivity anomaly beneath the epicentral zone.

3.3. VOLCANO-MAGNETIC EFFECT

Like the tectonomagnetic effect, the volcano-magnetic effect is also very complicated. For the volcanologists, the measurement of this parameter seems to be very important (as we have seen for seismic studies).

Measurements are fewer than for tectonomagnetic effects, mainly because the field experiments are difficult. Few results have appeared since 1965. At Mihara Volcano on Oshima Island, Japan, Yokoyama (1956, 1957) and Rikitake *et al.* (1963) found marked shifts in the orientation of the magnetic field a few months before eruptions. Before two eruptions of Mauna Loa, Hawaii, peculiar disturbances were observed in the vertical component of the magnetic field 200 miles away in Honolulu (MacDonald, 1951); but the cause of these disturbances or even whether they were actually related to the Mauna Loa eruption, is now known. More recent eruptions of Kilauea have not been accompanied by similar disturbances.

A more significant result has been obtained on Mt Ruapehu, New Zealand by Johnston and Stacey (1969). As one can see in Figure 11, significant variations were observed, mostly before eruptions. These results have been obtained by differential measurement techniques with 2 magnetometers placed about 8 km apart. The effects vary from one to several nT. However, it seems that this kind of magnetic effect does not appear systematically and the authors report that 'the magnetic preindication could be useful in at least some cases; more examples are needed for an assessment of their reliability'. More recent experiments have been made in New Zealand by Hurst and Christoffel (1973) on the White Island Volcano, yielding observations of modification of the magnetism of rocks by thermal processes. In the present case, the authors were optimistic, and concluded 'in volcanic centres like White Island, monitoring of both magnetic field and ground levels, taken in conjunction, show promise for predicting eruptions'.

This type of observation had also been made on Kadovar Island (Papua New Guinea) by Norris (1978) where he has shown a relation between hot vents and demagnetization.

The last experiment we should report is that which has been monitored over 'La Soufrière of Guadeloupe' volcano during the seismo-volcanic crisis of 1976–1977 by

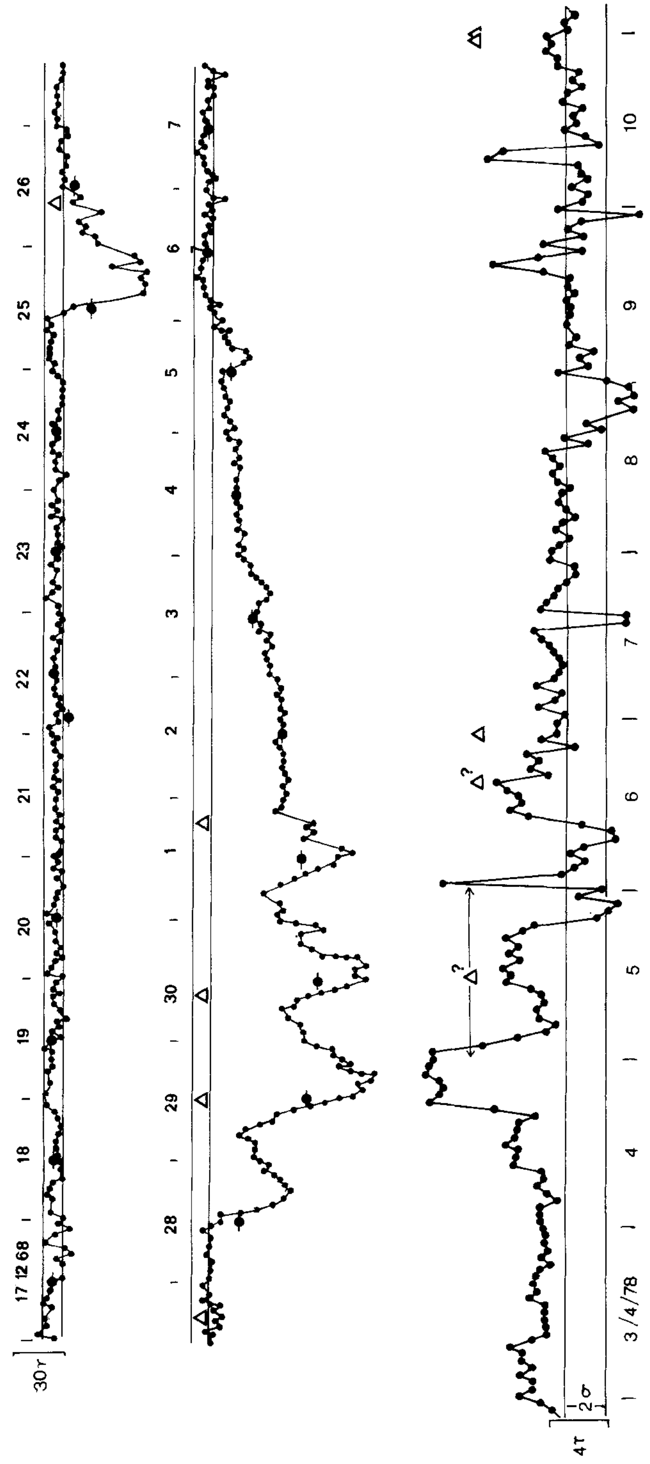

Fig. 11. Examples of volcano-magnetic effects observed at Mt Ruapehu (after Johnston and Stacey, 1969).

Pozzi *et al.* (1979).

An array of fourteen stations for measurement of the total intensity has been installed over the Soufrière massif; measurements were performed daily from September 1976 to April 1977. The main conclusions of these experiments are the following:

(1) Significant variations are found in the differences in intensity with characteristic time constants of a few days. These variations are very small for stations located near the summit and within 5 km of the reference station; they reach an amplitude of 15 nT for the more remote stations and often show good correlation from station to station (Figure 12).

(2) The amplitude of these variations decreased with the end of the seismovolcanic crisis at the end of March 1977.

From the few results related to volcanomagnetic effects, it appears that there are two kinds of mechanisms which can explain the observations.

First: thermal mechanisms, by heat transfer through modification of temperature by fumarolic activity and through the appearance of new fumarolic zones. The difference of temperature is involved in explaining the observed differences.

The second mechanism is more complicated because it does not in general exclude the first type. In this case, magnetic properties of the rocks are modified by pressure and so can be an effective method for monitoring explosive volcanism. As we shall discuss in the summary it may be an efficient method for forecasting.

4. Discussion

Small but significant *H* signals recovered from noisy magnetic records have tentatively been assigned to geodynamic manifestations like faulting or volcanic eruptions.

Two major problems appear which are generally not discussed. First, can the observed magnetic perturbation be reliably related to other phenomena such as ground deformation, tremors, etc., and are time-dependent parameters compatible?

Second, can the observations in one area be extrapolated to other situations on some justifiable basis?

It appears that the results obtained are greatly dependent on the mechanism involved, particularly regarding the time constants. These mechanisms are discussed in more detail by Beamish in his review.

There exist 'inefficient' mechanisms because of their long time-constant compared to the human scale or to the phenomena observed. A typical example of this kind of phenomenon is the exchange of heat between deep seated magma and the surface by conduction through rocks. This phenomenon may be useful in modelling the thermal state of the Earth and its evolution. It should be based on very long series of observations over large areas. The best way to study it is to map the magnetic field and to compare the maps at different epochs.

More interesting mechanisms are those which involve relatively short time constants and for which the integrative character of the propagation of the heat into the ground

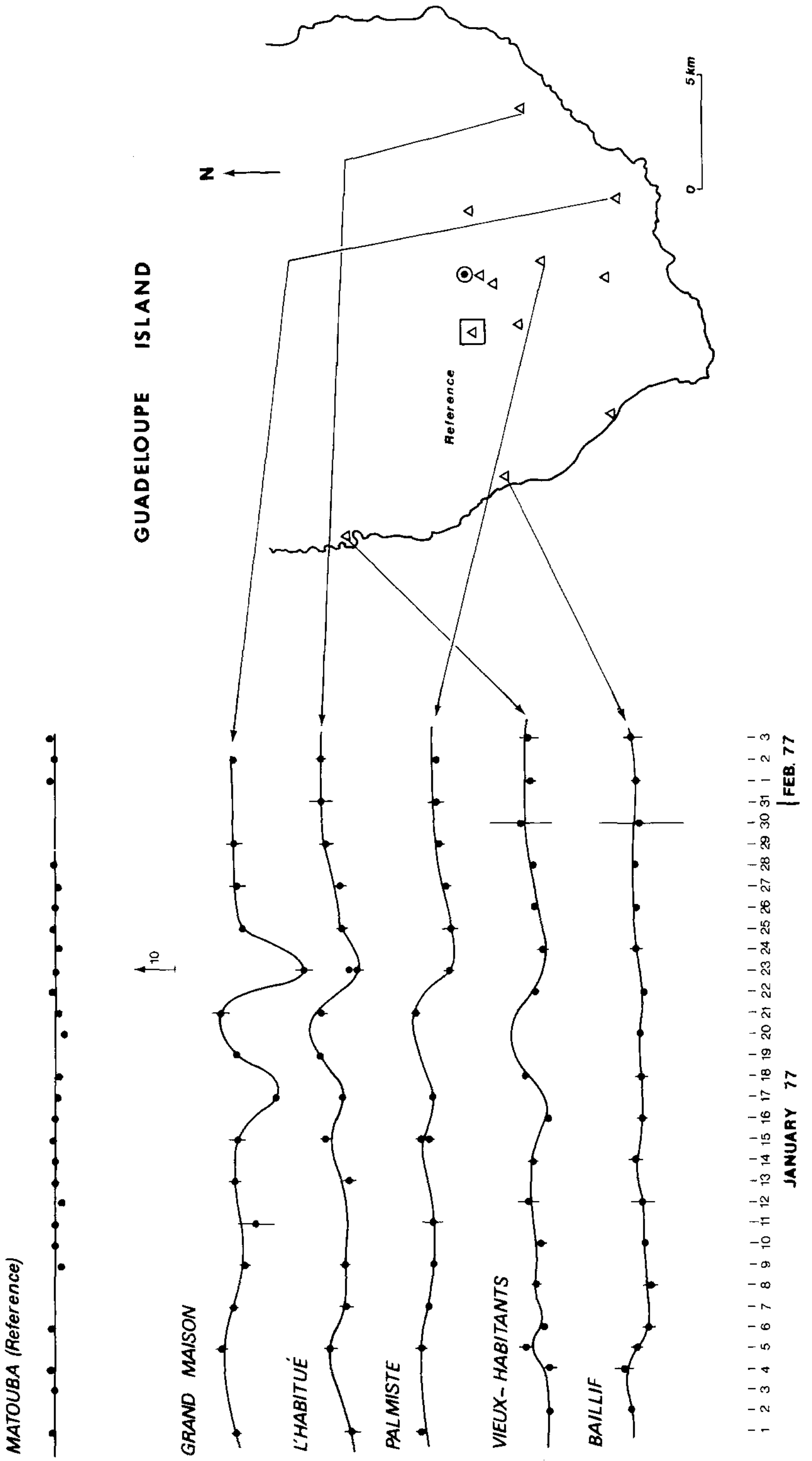

Fig. 12. Volcanomagnetic effects observed over the Soufrière volcano (after Pozzi *et al.*, 1979).

does not exist. Two major mechanisms of this kind are involved: stress effects (piezomagnetism and dilatancy) and electrokinetic.

Piezomagnetism depends only on stresses and the associated magnetic effects observed are largely reversible. There is no electric field associated with this phenomenon.

Dilatancy (the creation of cracks in stressed rocks) may also affect rock resistivity by modifying the pore structure and therefore influence the electrical conductivity (Brace *et al.*, 1965; Brace and Orange, 1968). Some models that take these effects into account are well known (e.g. Scholz *et al.*, 1973).

It seems possible to relate piezomagnetic effects and dilatancy by high pressure experiments such as those made by Pozzi and colleagues (personal communication).

The last kind of phenomenon that can explain the observed effect is electrokinetic filtration which corresponds to water and electrolytes flowing into pores. This phenomenon is well known but modelling in natural situations is too complex to be treated mathematically.

Whatever the kind of phenomenon involved, observations in one area are generally considered to be applicable to any other structure. Earthquake focal mechanism studies, even though not always very accurate, show that the strained volumes cannot induce important magnetic disturbances several hundred kilometers away.

For the volcano-magnetic observations, the problem is much more complex because the kind of predominant mechanism depends on the eruptive process involved and on the mechanical and physical properties of the volcanic edifices.

For the Kilauea volcano, which is the best monitored to-day, rocks are basaltic with high magnetic susceptibility, but magma does not generate explosions when it erupts. Pressure in the volcano edifice is not important. On the other hand, for volcanoes like the Soufrière of Guadeloupe or the Mount St-Helens in U.S.A. (Johnston, unpublished results), there are some important explosions and probably important high pressure into the volcano, even if rocks are poorer in magnetic minerals. The third case is Etna where the magma freely communicates with the atmosphere and it is very magnetic.

From these remarks, it is evident that all these observations have to be planned in conjunction with geologists, seismologists and geochemists and discussed in a global tectonic model. We have looked at the relation that can exist between the areas of the world presently studied and plate frontiers. Clearly, monitored areas are few. If we look at seismomagnetic studies, we find that these are restricted to transform faults (San Andreas fault and Turkey) or subduction zones (Japan). The situation is similar for volcano monitoring where well-studied volcanoes are essentially permanently active ones and so less dangerous, and those where intrapressure is not important; one exception has to be made: the Soufrière of Guadeloupe.

Even if the magnetic investigation in Geodynamics is not well coordinated at the international level, it seems that these effects can be a very important tool for forcasting because, theoretically, effects are to be expected before the geophysical events (earthquakes and volcanic eruptions) occur.

4.1. SUMMARY

The introduction of proton magnetometers for the measurement of the magnetic field and the continuous differential monitoring at two or more stations, has permitted differences between stations to be obtained with an accuracy better than 1 nT. The differential technique will yield time-averaged signals of less than 1 nT which is usually greater than the variations associated with current-channelling in anomalously conductive structures. The channelling effects may generally be accounted for through a detailed study of normal transient variations (e.g. Pozzi *et al.*, 1979) in restricted zones. It is to be hoped that such data reduction techniques may permit the isolation of signals of 1–2 nT amplitude. Experimental and theoretical modelling of the volcano and seismo-magnetic effect results in order-of-magnitude agreement with observations (Rikitake, 1976).

Important developments should be instigated in data reduction of continuous magnetic field measurements. Future research should be focused on several new sites for the application of this technique and the relationships between different geodynamic phenomena and their magnetic effects (e.g. eruptive mechanisms and rock types for volcano-magnetism; focal mechanism and magnitude for earthquakes).

Finally, studies of the time-constants associated with different natural causes for tectono-magnetic effects should precede the use of geomagnetic field monitoring as a forecasting technique.

4.2. THE USE OF MAGNETO-GEODYNAMIC EFFECTS FOR HAZARD FORECASTING

In conclusion, we should like to present one diagram (Figure 13) which is an attempt at

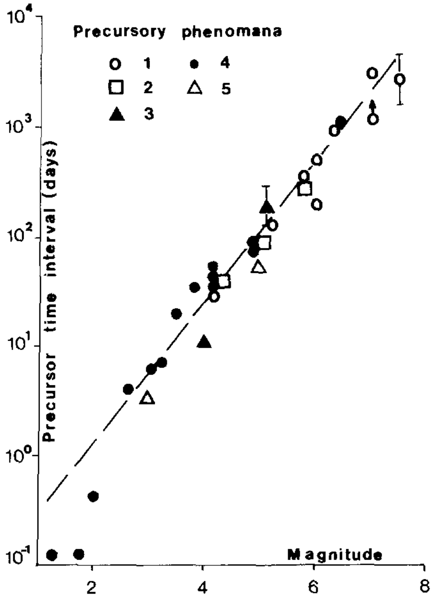

Fig. 13. Duration time of various precursor phenomena as a function of earthquake magnitude. (1) crustal movements; (2) electrical resistivity; (3) radon emission; (4) Vp/Vs anomaly; (5) b values. (from Rikitake 1979).

using the seismomagnetic effect for forecasting. On this curve, it appears that magnetic events are produced before earthquakes occur and that the time-constant depends on its magnitude. This kind of relationship, if confirmed, could be used for hazard assessment if not for forecasting.

Identification of a magnetic signal as a precursor to a geophysically hazardous phenomenon appears to be a promising technique for forecasting. The day and place where it could be applied successfully may not be restricted to science-fiction.

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