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On Possibilities to Extract Scientific Information from Magnetic Field Recordings on the Martian Surface – Global/Regional Viewpoint

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Abstract

This paper deals with possibilities of interpreting magnetic data recorded at one or more fixed stations on the surface of Mars. The discussion is to a great extent based on studies about the terrestrial magnetic field. The static main field and the time variations are considered separately. The former is composed of contributions of different spatial scales, and the latter is the sum of the fields due to ionospheric–magnetospheric sources and currents induced within the ground. A decomposition of the different contributions is difficult in practice, which restricts the possibilities of getting geophysical information by magnetic measurements. We conclude that the number of three-axis magnetometers operating on the Martian surface should be at least from five to ten to lead to scientifically firm and quantitative results. Of course, even one magnetometer may yield for example statistical information about the temporal behaviour of the magnetic field.

1. *Introduction*

Magnetic studies on a planetary surface can basically be divided into two types:

i) "Microscopic" investigations in which soil samples are analyzed physically and/or chemically to reveal magnetic properties of the material.

ii) "Macroscopic" investigations in which a magnetometer is recording the magnetic field.

This paper deals with the latter topic.

Several missions are being planned for scientific exploration of Mars in the next ten years: e.g. Mars–94/96 (Russia), Mars–98 (Russia), Mars Surveyor Program (USA) and InterMarsnet (ESA); for references, see *INTERMARSNET* (1993), *MARS–94* (1992), *Siili and Pirjola* (1994) and *IMEWG* (1994). Landers to make measurements on the surface are also included in these plans, and one of the instruments is a magnetometer. Our purpose is to give an overview of interpretation possibilities of magnetometer data. Thus, we assume that the magnetic field is recorded at one or more fixed stations on the surface of Mars. The absolute value of the three–component vector field, whose direction is fixed with respect to Martian geographical coordinates, is recorded at Martian sites at suitable time intervals (10 s to 1 min) during a long time period (≈ 1 Martian year). The accuracy of the data will be ±0.1 nT. Because the physics of the Martian magnetic environment is quite poorly known, references and terminology principally concern studies performed on the Earth.

The possible magnetic field of a celestial body can be divided into two parts:

1) Main field varying so slowly that it can be regarded as static in time scales less than a few years, and this kind of a secular variation is not considered in this paper.
2) Temporally rapidly varying part with time scales from some seconds to one day (note: one Martian day ≈ 24 h).

The main field further consists of three parts:

1) Intrinsic global field originating deep from the interior of the planet (from the possible liquid core and dynamo processes therein). A typical length scale is some thousands of kilometers.
2) Field produced by regional anomalies at different depths within the planet. Length scales are typically in the order of tens to hundreds (or even thousands) of kilometers.
3) Small–scale field due to local magnetized rocks, boulders and other material. Typical scales are from some metres to a kilometer.

It should be noted that this is the definition of the "main field" in this paper; in other connections the mere intrinsic field can constitute the "main field".

Part 3 above is not discussed in this paper, because the magnetometers are assumed to be at fixed locations and distant from each other (≥ 100 km). So rocks and boulders are difficult to be detected; for references *Kukkonen et al.* (1991) and *Terho et al.* (1993) can be mentioned.

The temporally–varying magnetic field is composed of two parts:

1) Primary contribution of ionospheric–magnetospheric currents.
2) Secondary field created by induced currents within the ground (telluric currents).

(Here the "ionosphere" and "magnetosphere" of a celestial body refer to its near–space also in case that these "spheres" do not exist in a way analogous to the Earth.)

Studies of each of the five parts of a planetary magnetic field are scientifically useful and valuable, and any of them might be a scientific objective of a mission. In general, the main field and the variation can be treated separately. However, the main field has an indirect influence on the variation field, because the interaction between the solar wind and the main field determines the structure of the planet's ionosphere and magnetosphere.

The specific features of the Martian magnetic field are outlined in Section 2. The main field and the variation field are discussed in Sections 3 and 4, respectively. Section

5 finally summarizes our conclusions about possibilities of scientific interpretations of magnetic measurements on Mars.

2. *About the Martian geomagnetic environment*

Until now, magnetic measurements have been performed in the near–space of Mars, but not on the surface. A great part of our present knowledge about the Martian magnetic environment is based on the Soviet Phobos–2 mission at the end of the 1980's (*Zakharov*, 1992; *Kallio*, 1992). However, the question of the possible existence of an intrinsic magnetic field produced by electric currents deep within the planet is not yet solved. It is only known that the field is very weak having values in the order of tens of nT or less at the surface. Such a magnitude is clearly less than that predicted by the magnetic Bode's law (*Russell*, 1980; *Merrill and McElhinny*, 1983, p. 347; *Chiappini and De Santis*, 1994). The law is empirical, and simply states that the magnetic moment of a celestial body is logarithmically proportional to the angular momentum. It is evident that there exists no active dynamo, which would produce a magnetic field, in the interior of Mars (at least any more) although the planet probably has a core (*Russell*, 1980; *Merrill and McElhinny*, 1983, p. 350; *Chiappini and De Santis*, 1994).

A Martian magnetic field of a few tens of nT is close to the limit where the field is strong enough to create a pressure balance against the solar wind thus producing an Earth–like magnetosphere (*Intriligator and Smith*, 1979; *Kallio*, 1992). Consequently, an accurate measurement of the magnetic field is of utmost importance. If a Martian magnetosphere can be defined at all, its size is anyway small compared to the Earth's magnetosphere; the altitude of the magnetopause at the subsolar point is in the order of hundreds to a thousand kilometres (*Breus*, 1992; *Kallio*, 1992).

Observations made by the two US Viking landers in the end of the 1970's show that there is magnetic surface material on Mars (*Hargraves et al.*, 1979). SNC meteorites most probably of Martian origin and found on the Earth (*Laul*, 1986) indicate the existence of remanently–magnetized rocks and boulders at the surface of Mars. A theoretical estimate based on measured properties of SNC meteorites suggests local magnetic anomalies of the order of even several thousands of nT in the vicinity of magnetized boulders (*Terho et al.*, 1993). Magnetization induced by the weak Martian field in materials with moderate susceptibilities may be considered insignificant.

3. *Main field*

3.1 *General*

The sources of the static main field defined in Section 1 are located at or below the surface of the planet. A naturally–looking possibility to get the value of the main field at a station is to average the data over a sufficiently long time interval, say some days.

However, it is not clear whether the average of the contribution from ionospheric–magnetospheric sources is zero, and even an influence of the interplanetary magnetic field may remain. This can be seen by considering for example a ring current which may cause a "variation" field not averaging to zero. Nevertheless, the static field so obtained is representative of the particular site, and if a spherical harmonic analysis of the magnetic field can be performed, information about magnitudes of contributions from the planet's interior and exterior is received (see Section 3.2, and e.g. *Chapman and Bartels*, 1940, pp. 606–698; *Merrill and McElhinny*, 1983, pp. 17–31; *Langel*, 1987).

3.2 Global view

It is essential to note that the main field observed on a planetary surface only represents locally the particular site, and a generalization to the global magnetic field is quite ambiguous.

However, due to different locations and sizes of the sources, small–scale, regional and global parts of the main field "look" magnetically different when the lander is approaching the surface of the planet. Consequently, descent–phase magnetic data can help in the separation of contributions of different spatial scales. On the other hand, ionospheric–magnetospheric currents also affect the data distorting the interpretation. Thus, a single station on the surface cannot yield much global elucidation of the main field.

To obtain a quantitative global understanding of the main field, a spherical harmonic analysis (see Section 3.1) should be carried out based on data collected at more than, say, ten stations covering the whole planet. The analysis will show if the field can satisfactorily be described by a dipole lying at the centre of the planet as is the case on the Earth. The type of the main field is important regarding the interaction with the solar wind. Even ten is quite a small number of stations, since for example Gauss used 84 data points when analysing the terrestrial field in 1838 (*Chapman and Bartels*, 1940, p. 640), but about ten might be sufficient for a rough estimate of the nature of the field. If the secular variation can be considered insignificant, the magnetometers measuring a planetary main field need not operate simultaneously, but data to be collected for example at later Martian missions can be combined with Mars–94/96 Small Station and Penetrator magnetic recordings to be obtained in the nearer future (*Pirjola et al.*, 1994).

Of course, by making an assumption that the field has a dipole form, the strength and direction of the dipole located at the planet's centre can be calculated using the vector magnetic field at a single station on the surface. Analogously, using several stations makes it possible to determine a dipole, a quadrupole, and higher multipoles depending on the exact number of datapoints available. However, this kind of an analysis only produces equivalent multipoles valid for the particular sites and need not have to do much with reality at other parts of the planet (and cannot be termed "spherical harmonic analysis").

Thus, determination of the main field in an unambiguous global manner based on measurements at fixed surface stations is a difficult and hard task. Evidently, orbiter observations of a possible dipole–like intrinsic field could be much more useful. However, the field due to ionospheric–magnetospheric currents may dominate over the intrinsic field at orbiter heights.

3.3 Regional anomalies

Regional anomalies distort the global field. At least in the case of the Earth, their sources lie near the surface in the crust or are local disturbances in the current system on the surface of the liquid core of the planet. Two or more stations at distances in the order of tens to several hundreds of kilometres might reveal such anomalies. As indicated in Section 3.2, regional anomalies can also be observed by analyzing the descent–phase data. Besides, a magnetometer installed on an orbiter "sees" regional anomalies differently from the global field.

3.4 Magnetized rocks and boulders

Small–scale anomalies created by local magnetized rocks or boulders at the magnetometer may completely hide other parts of the main field. However, utilizing for example camera observations may provide information about the possible existence of magnetized rocks and boulders, and about their shapes and locations, so their influence on the magnetic field can be estimated.

Using another magnetometer (or several) at a distance from tens of metres to one kilometre permits the investigation of small–scale anomalies, and descent–phase data can be useful, too (see Section 3.2). A study of magnetized rocks and boulders would be much more effective, and thus more recommendable, by a rover moving on the surface of the planet (e.g. *Kukkonen et al.*, 1991).

Analyses of the magnetism of rocks and boulders is scientifically significant since, besides yielding information about the structure and composition of the ground of the planet, the magnetization may also contain paleomagnetic information about the ancient magnetic field of the planet and is thus directly associated with the planetary evolution history.

4. Time varying field

4.1 General

Ionospheric–magnetospheric currents are the primary reason for the temporally and spatially rapidly varying magnetic field. The space currents are in turn influenced by the interaction between the solar wind and the main magnetic field. Another possible mechanism producing ionospheric currents is the tidal effect of the atmosphere.

The ionospheric–magnetospheric currents produce the primary (external) magnetic field and an associated electric field. The primary electric field in turn induces (Ohmic) currents in the conducting ground, and they cause the secondary (internal) electromagnetic field. The total magnetic variation observed on the ground is a sum of these two contributions. The purpose of a research of magnetic variations at a planetary surface can be the investigation of ionospheric–magnetospheric currents or the determination of the conductivity structure of the ground.

A theoretical treatment of the variation field is based on the Maxwell equations. If the primary sources and the ground's electromagnetic structure are known, the electromagnetic field can be calculated. In practice the main task is an inversion problem: resolve the sources and/or the electromagnetic structure of the ground when the electromagnetic field is measured. In some cases, additional data like electric field recordings on the surface or in the ionosphere, or geological knowledge may be available.

4.2 *About resolution of inversion*

Before going to details, we briefly discuss the length scales of the structures that can be detected using magnetometer data. The following lengths are important:

1) Height h of the ionospheric currents (about 100–120 km for the Earth, unknown for Mars).
2) Typical distance d between magnetometer stations.
3) Skin depth δ in a (homogeneous) ground, where

$$\delta = \frac{2}{\sqrt{\omega\mu\sigma}} \tag{1}$$

Here ω is the angular frequency, μ is the ground's permeability and σ is the ground's conductivity. Note that δ can be exactly defined only for a plane wave in a homogeneous medium. The skin depth in earth materials is illustrated in Fig. 1.

The first two lengths restrict the spatial resolution of the inversion of ionospheric–magnetospheric currents. Roughly stating, spatial details smaller than the maximum of h and d cannot be observed using ground–based recordings. So, if the aim is to study ionospheric currents, d is reasonable to be chosen to approximately equal h.

The study of the ground conductivity is different in the sense that the smaller d the smaller details can be detected. However, the skin depth roughly dictates the depth below which information is not available.

Typical distances between magnetometers in routinely working arrays, like IMAGE operating in Fennoscandia now, are 200–300 km (*IMAGE Newsletter No. 1*, 1992). For shorter measuring campaigns, like IMS (International Magnetospheric Study) in Fennoscandia in the end of the 1970's (*Küppers et al.*, 1979), a denser network (100–150 km) was used. Even shorter distances of only some tens of km have been used for magnetovariational and magnetotelluric soundings (e.g. *Pajunpää*, 1984; *Korja et al.*, 1986).

The time scales of magnetic variations must be known to choose the sampling interval of the recordings. As the first approximation, we believe that magnetic variations on other planets have similar temporal characteristics as terrestrial variations because both are essentially affected by the solar wind. On the other hand, the conditions of the plasma environments of different planets vary very much, so their characteristic response times may also differ essentially. Anyway, the same sampling intervals, typically from about ten seconds to a minute, applied on the Earth should be used elsewhere, too. The sampling intervals also affect the frequencies involved (cf. Nyqvist frequency) and thus limit the depth ranges that can be investigated within the planetary interior with electromagnetic methods (cf. Eq. (1)).

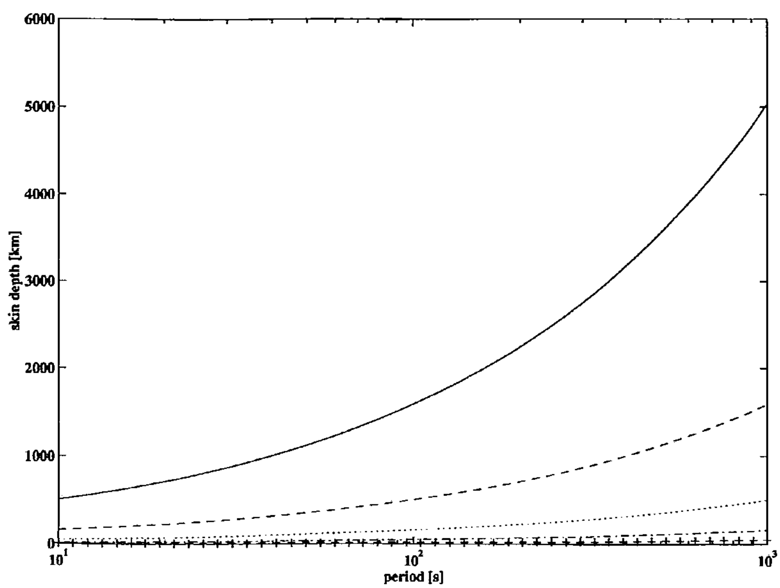

Fig. 1. Skin depths of five media having conductivities 10^{-5} (–), 10^{-4}(- -), 10^{-3} (:), 10^{-2} (–.) and 10^{-1} $\Omega^{-1}m^{-1}$ (+). The values correspond to earth materials, and the periods shown refer to time–scales of typical geomagnetic variations. The permeability has the vacuum value.

4.3 *Separation of the variation field into external and internal parts*

The main problem in the use of magnetic variation data concerns the possibilities of separating the external and internal contributions in a recorded magnetic variation. It is difficult even on the Earth where much more is known about the geoelectromagnetic environment than on other planets, measurements can be repeated, recording sites may be changed, a dense array of stations can be created, there are no limitations of mass and power, etc.

If the temporally varying field is known everywhere at the planet's surface, it can be uniquely separated into external (due to ionospheric–magnetospheric currents) and internal parts (due to telluric currents). This holds true for both planar geometry suitable for regional studies, and for spherical geometry based on spherical harmonic analysis and suitable for global studies (*Berdichevsky and Zhdanov*, 1984). Examples of practical applications in areas of the order of 1000×1000 km^2 are given by *Porath et al.* (1970), *Mersmann et al.* (1979) and *Richmond and Baumjohann* (1984). Such investigations require 2–D magnetometer arrays with at least about ten stations.

A magnetometer chain with a smaller number of stations (roughly five or more) may also be used for the separation assuming that

1) Ionospheric currents flow only into one direction.
2) Currents do not vary spatially in the flow direction.
3) Ground conductivity only depends on depth. (As a special case, it may also depend on the horizontal coordinate transverse to the current flow.)

Magnetometer chains located under such ionospheric currents systems have been used (e.g. *Küppers et al.*, 1979).

If both the ionospheric–magnetospheric currents and the structure of the ground are poorly known, a mathematical separation outlined in this section seems to be the only reliable way to extract geophysical information from magnetic variation data.

4.4 Inversion of surface magnetometer data into ionospheric–magnetospheric currents

If the external part of the surface field is known, a source–free horizontal ionospheric equivalent current distribution can be determined that explains the ground recordings. However, there may also be an irrotational current system together with field–aligned currents whose contribution cannot be detected on the ground. If additional data (e.g. the electric field in the ionosphere and/or the ionospheric conductivities) are available, there are sophisticated methods for determining also the irrotational and field–aligned current system (e.g. *Glassmeier*, 1987). An example of the upward continuation of a two–dimensional ground magnetic field is given by *Mersmann et al.* (1979) with data from seven stations.

A direct modelling with free parameters can also be used. Simplifying assumptions about the conductivity structure of the ground help the treatment; for example, the conductivity may be so small that induction is negligible as the first approximation, or the conductivity structure can be described by a perfect conductor at a given depth. With such assumptions, which could be based for example on seismology, or on models obtained from observations of static magnetic anomalies, ionospheric–magnetospheric currents may be successfully studied using about five stations. As indicated in Section 4.2, favourable distances between the stations are dictated by the height of the currents. Five stations with such distances do, of course, not give a global picture of the current systems in the near–space of the planet, but it can yield information of currents flowing in a specific area, similarly to studies of the terrestrial auroral electrojet.

An example of the modelling of a 2–D current flow is given by *Lühr et al.* (1994) who used five EISCAT magnetometer stations to determine the total ionospheric current and its transverse distribution; EISCAT magnetometers were the predecessor of IMAGE mentioned in Section 4.2. A 3–D modelling of ionospheric currents based merely on magnetic recordings has been applied e.g. by *Bannister and Gough* (1977) who used a Canadian array of 25 magnetometers. A similar technique with simultaneous observations of the ionospheric electric field has been applied e.g. by *Baumjohann et al.* (1981) who used the Scandinavian IMS magnetometer data of about 25 stations. It must be noted that for example additional electric measurements in the ionosphere reduce the ambiguity of the current inversion.

4.5 *Inversion of surface magnetometer data into the ground conductivity*

Even without a separation into external and internal contributions, a study of the conductivity structure of the ground is possible provided assumptions of the ionospheric–magnetospheric currents may be made. The latter could be based for example on data collected by orbiters. On the Earth the usual (but not always valid) "plane wave" assumption is that the field produced by space currents is horizontally homogeneous, which is the case if the field is produced by a wide uniform current sheet or by a distant source. However, if the plane wave assumption is not well–established, the geophysical interpretation of the results becomes random and completely incorrect conclusions may be drawn. This is the source effect that is often present at auroral latitudes of the Earth (e.g. *Mareschal*, 1986; *Osipova et al.*, 1989).

4.5.1 Geomagnetic induction vectors

Single station data gives a possibility for a qualitative study of conductivity anomalies by calculating geomagnetic induction vectors (e.g. *Schmucker*, 1970). Their definition is based on transfer functions $A(\omega)$ and $B(\omega)$ between the vertical (B_z) and horizontal magnetic field components (B_x and B_y) in the frequency domain:

$$B_z(\omega) = A(\omega)B_x(\omega) + B(\omega)B_y(\omega) \tag{2}$$

The real and imaginary induction vectors are usually defined by the formulas

$$\mathrm{S}_{\mathrm{re}} = -\mathrm{Re}(A(\omega))\mathbf{e}_x - \mathrm{Re}(B(\omega))\mathbf{e}_y \tag{3}$$

$$\mathbf{S}_{\mathrm{im}} = \mathrm{Im}(A(\omega))\mathbf{e}_x + \mathrm{Im}(B(\omega))\mathbf{e}_y \tag{4}$$

The interpretation of the real vector is that it points towards current concentrations within the ground, i.e. approximately towards highly–conducting regions. The imaginary vector is obviously related to near–surface structures, but its interpretation is not straightforward.

As mentioned above, the basic assumption for these interpretations is that the primary variation field is a plane wave. If this is not the case, induction vectors may characterize ionospheric–magnetospheric currents rather than the ground structure.

Even if the plane wave assumption is valid, the induction vectors can give only qualitative information of the location of conductivity anomalies. For quantitative analysis, either additional recordings of the electric field (magnetotelluric sounding) or a wide magnetometer array are needed (Sect. 4.5.2).

4.5.2 Method of horizontal spatial gradients

Magnetotellurics (MT) based on simultaneous electric and magnetic recordings at a single station on the surface constitute a method widely used in studies of the Earth's structure. The basic quantity is the surface impedance

$$Z = \frac{\mu_o E_x}{B_y} \tag{5}$$

where E_x and B_y are perpendicular horizontal electric and magnetic field components on the surface, respectively, and μ_o is the permeability in free space. In principle, if the conductivity structure is layered, the conductivity can be uniquely inverted using MT data only from one station (*Weidelt*, 1972). However, even in such a simple case the inversion is mathematically an improperly posed problem, where small changes in the data can cause large changes in the results. For 2–D and 3–D structures the problem is still more difficult.

The magnetotelluric surface impedance can also be calculated from mere magnetic data if the horizontal spatial derivatives of the magnetic field are known. It can be shown that in the frequency–wave number domain the impedance

$$Z_g = i\omega\mu_o \frac{B_z}{\dfrac{\partial B_x}{\partial x} + \dfrac{\partial B_y}{\partial y}} \tag{6}$$

yields the same information of the ground conductivity as Z (e.g. *Schmucker*, 1970). Now it is essential that the "horizontal spatial gradient" $\dfrac{\partial B_x}{\partial x} + \dfrac{\partial B_y}{\partial y}$ can be determined carefully, which requires a two–dimensional and dense enough magnetometer array. An example of such an approach is given by *Jones* (1980), who used 10 of the Scandinavian IMS magnetometers. This number has to be considered the minimum for a reliable calculation of the derivative term.

Alternatively, we may have only one or a few magnetometers and presume the spatial behaviour of the field (or the characteristic wave numbers), for example by assuming the ground structure as one–dimensional and utilizing orbiter data about ionospheric–magnetospheric currents. Such an investigation can, however, be very ambiguous, and the results cannot be considered unique.

It should further be noted that even if the magnetotelluric surface impedance is calculated correctly its interpretation is extremely difficult unless the "plane wave" assumption of the ionospheric–magnetospheric contribution is at least roughly satisfied (e.g. *Pirjola*, 1992; *Viljanen et al.*, 1993). In Fig. 2, we show the "apparent resistivity" (essentially the square of Z in Eq. (5)) in a case where the ionospheric source is a sheet current. In that example, the ground is homogeneous with a resistivity of 1000 Ωm. The apparent resistivity gives the equivalent constant resistivity for any ground structure, and in the case of a homogeneous ground it equals the true resistivity.

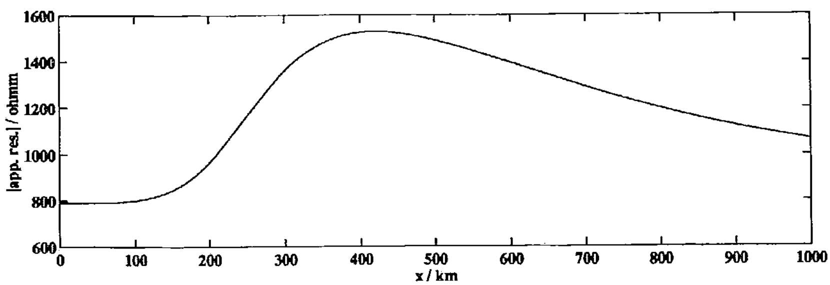

Fig. 2. Apparent resistivity due to an infinitely long uniform sheet current of a width of 400 km flowing a height of 110 km. The centre of the electrojet is at x=0 km. The earth is homogeneous and has a conductivity $10^{-3}\ \Omega^{-1}\text{m}^{-1}$. The period of the time–harmonic variation is 300 s. For a plane wave, the apparent resistivity is 1000 Ωm, and it is independent of x.

In MT studies on the Earth the magnetic susceptibility may usually be neglected, but this need not be the case for other planets. Thus, ignoring the possibility of susceptible material can lead to incorrect conclusions of the conductivity structure. It is illustrated in Fig. 3. Two completely different fictitious two–layer models are considered (Fig. 3a): one with a thin magnetically susceptible upper layer and the other with a thicker non–susceptible and less–conducting layer. Fig. 3b shows that the two structures give very similar plane wave apparent resistivities as functions of the period.

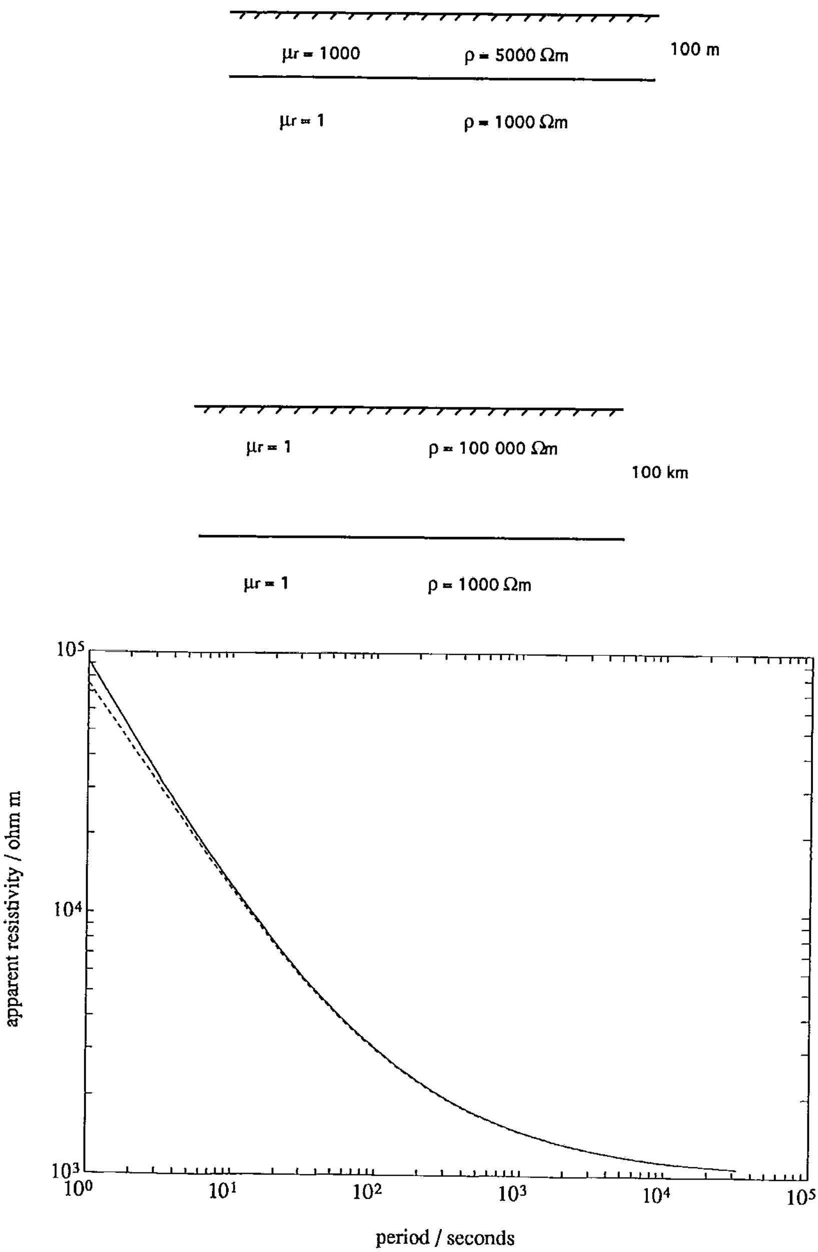

Fig. 3. a) Two different two–layer ground structures.
b) Plane wave apparent resistivities as functions of period for the two structures. Solid line corresponds to the upper model, and dashed line to the lower model, respectively.

5. *Conclusions*

Because the amount of data of the Martian magnetic field is very small so far, all new magnetic recordings on the surface of the planet would in principle be welcome. However, both the main field and the temporal variations contain several parts having different origin, so an interpretation of the data is difficult.

A sufficiently long time series of magnetic recordings with short enough sampling intervals even at a single surface station permits investigations of the morphology and characteristic frequencies of Martian magnetic variations, and statistical conclusions can be drawn. Recommendable lifetimes, sampling intervals and accuracies of magnetometers operating on the Martian surface should be at least a Martian year (about two Earth years), 10 s to 1 min, and 0.1 nT, respectively, and three–component data with absolute zero-levels should be collected. Additional data from an orbiter or from the lander when descending towards the surface of the planet could essentially supplement such results, but the conclusions would certainly remain quite poor and concern only the recording area.

Single–station magnetic variation data also permits the computation of geomagnetic induction vectors usable in studies of the structure of the planetary interior, but to lead to reliable results, the field produced by space currents should be horizontally homogeneous (i.e. a "plane wave"). The latter information might be provided by orbiter observations.

Nevertheless, we regard magnetic measurements at fixed stations on the surface of Mars as extremely difficult and ambiguous to be interpreted unless the number of stations is about five to ten or more. This conclusion concerns investigations of both the main and the variation field. The stations could be distributed all around the planet to yield global information or in a smaller area to be used for regional studies. Additional data for studies of the spatial structure of the magnetic field can be obtained by an orbiter.

The criticism against magnetometers is further supported by the fact that magnetic measurements are apt to be distorted by magnetic fields created by the station itself. This is a serious electromagnetic compatibility problem, and to avoid it, special laborious procedures are required (e.g. *Pirjola et al.*, 1994).

Of course, all the matters discussed in this paper can easily be tested using magnetic data collected at observatories on the Earth: First use data of a single observatory and try to estimate the characteristics of the main field, the ionospheric–magnetospheric currents and the Earth's structure. Then use additionally recordings of another observatory and/or data obtained by an orbiting satellite, and so on.

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