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Magnetism, Iron Minerals, and Life on Mars

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ABSTRACT

A short critical review is provided on two questions linking magnetism and possible early life on Mars: (1) Did Mars have an Earth-like internal magnetic field, and, if so, during which period and was it a requisite for life? (2) Is there a connection between iron minerals in the martian regolith and life? We also discuss the possible astrobiological implications of magnetic measurements at the surface of Mars using two proposed instruments. A magnetic remanence device based on magnetic field measurements can be used to identify Noachian age rocks and lightning impacts. A contact magnetic susceptibility probe can be used to investigate weathering rinds on martian rocks and identify meteorites among the small regolith rocks. Both materials are considered possible specific niches for microorganisms and, thus, potential astrobiological targets. Experimental results on analogues are presented to support the suitability of such *in situ* measurements. **Key Words: Early life—Mars—Magnetism—Iron minerals.** Astrobiology 6, 423–436.

INTRODUCTION

In the last decade, meteorites (SNCs) and *in situ* exploration have been used to study the magnetic field of Mars and the magnetic properties of martian materials. Although studies have mostly focused on the internal dynamics of Mars, results also have implications for the putative appearance and development of martian life and the habitability of the planet.

Magnetic fields are created by electric currents (in the planetary core, ionosphere, lightning channel, or solar wind) or by magnetization of matter. Magnetization can either be induced by the ambient field or be remanent [natural remanent magnetization (NRM)]. NRM provides information on past magnetic fields in which the material was magnetized. The study of NRM defines the discipline of paleomagnetism (Dunlop and Ozdemir, 1997). Magnetization of natural materials is mostly linked to iron-rich minerals. Iron, due to its various oxidation states, is a key element in redox equilibrium on planetary surfaces with or without biological activity. The most prominent feature of Mars, noted since antiquity, is its reddish color due to the presence of Fe^{III} minerals. A key issue regarding the surface history of Mars is that of the processes, possibly bi-

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ologically mediated, responsible for the oxidation of primary forms of more reduced iron.

The present contribution first provides a short review of martian magnetism-related subjects from an astrobiology perspective. It then demonstrates the astrobiological interest of *in situ* magnetic measurements during future rover exploration of Mars. Measurement concepts will be illustrated by the example of the MAPPA instrument suite that was proposed for the 2009 NASA Mars Science Laboratory (MSL) mission. Although the proposed suite was not selected, it could provide a basis for future rover missions. The suite consists of two instruments that measure *in situ* objects: a contact magnetic susceptibility probe (COMASP) and a deployable magnetic remanence device (MAREDD). The detection depth or size scales of these probes are in the 1–50 mm and 5–100 cm ranges, respectively. A previous publication already put forward this concept (Rochette *et al*., 2004a) and contains technical details on the proposed experiments. However, this publication did not refer specifically to Mars and to astrobiological objectives. Other options for magnetic measurements could be put forward as well, and it is not our purpose here to discuss at length the merits of the various options. In particular, instruments measuring grabbed samples could also be envisioned.

HISTORY OF THE MARTIAN MAGNETIC FIELD

Until 1997, Mars was described as a non-magnetic planet, *i.e.*, devoid of an Earth-like global magnetic field of internal origin. The major discovery of the Mars Global Surveyor orbiter was the presence of intense local magnetic anomalies of up to a few microtesla produced by the NRM of crustal rocks (Acuña *et al*., 1999). These anomalies, with wavelengths of several hundreds of kilometers, are one order of magnitude smaller than the main Earth field generated by the dynamo effect in the core, but two orders of magnitude larger than typical crustal magnetic anomalies on Earth. Although the origin of these anomalies is largely unknown, they undoubtedly indicate the presence of a large Earth-like internal field (10–100 μT range) during NRM acquisition. Based on the location of the anomalies, which occur primarily on the heavily cratered older crust (Noachian age) of the southern hemisphere, Acuña *et al*. (1999) concluded that the internal field of Mars had a limited lifetime and must have shut down at about 4 Ga. However, there was some debate over this short lifetime (*e.g.*, Schubert *et al*., 2000). In particular, some weaker magnetic anomalies are also visible in the younger northern hemisphere, even in the youngest volcanic centers (Purucker *et al*., 2000). In the Noachian crust, the magnetization age is not the formation age of the surface (dated by cratering density) but the cooling age of the lower crust, which may be significantly younger. Nevertheless, more recent comprehensive studies on martian crustal magnetization distribution (Langlais *et al*., 2004) and age relations (Arkani-Hamed, 2004) do support a restriction of dynamo action in the Noachian, *i.e.*, no younger than 3.8 Ga. The major argument is linked to the fact that large impact craters do not show a large magnetization (above cited references as well as Hood *et al*., 2003). This implies that the dynamo had already shut down at the time of impact. The details of the impact demagnetization process are still a matter of debate (Rochette *et al*., 2003a, 2004b; Mohit and Arkani-Hamed, 2004). Hood *et al*. (2005) studied the magnetization direction of strong localized anomalies and hypothesized that the Noachian pole of rotation shifted approximately 60° since the time of magnetization. Furthermore, they observed that the proposed Noachian paleoequator, more or less, corresponds to the belt where both valley networks and large magnetization are concentrated. This correlation invokes an interesting perspective on interactions among climate, surface morphology, and crustal processes (see discussion in Rochette, 2006).

Paleomagnetic evidence from SNCs has been cited to support the discussion on the history of the dynamo. In terms of the standard interpretation of isotopic data (McSween, 2002), the magmatic ages of all SNCs, except ALH84001, are much younger than the late Noachian dynamo shutdown. Accordingly, they yield an estimated paleomagnetic field intensity of around 5 μT, *i.e.*, lower than an Earth-like dynamo and in the predicted range for the present surface field (Cisowsky, 1986; Collinson, 1997; Gattacceca and Rochette, 2004). On the other hand, Bouvier *et al*. (2005) challenged the standard interpretation and proposed that crystallization of all SNCs occurred around 4 Ga. However, as all SNCs have been shocked, the recorded field may date from the impact rather than crystallization (Cisowsky and Fuller, 1978; Rochette *et al*., 2003a), thus indicating a younger paleomagnetic age. Indeed, the Nakhlites, sometimes referred to as "unshocked" (see Shuster and Weiss, 2005), show clear evidence of strong shock [estimated pressure of 20 GPa and high temperature (Greshake, 1998; Malavergne *et al*., 2001)]. These conflicting views could be resolved considering the heterogeneous pressure–temperature conditions generated at the microscopic level during shock wave propagation in granular material (Baer, 2002). A further complication in the interpretation of paleomagnetic data in shocked rocks is the suggestion that transient fields due to impact-generated electric currents can overcome the steady ambient field during magnetization acquisition (Enemoto and Zhang, 1998; Carpozen *et al*., 2005; Soloviev and Sweeney, 2005).

Various estimates have been published for the paleointensity recorded by ALH84001. Weiss *et al*. (2002) presented evidence for a "high intensity" paleomagnetic field (with values in the 5–50 μT range), while Antretter *et al*. (2003) and Gattacceca and Rochette (2004) reported 5 and 20 μT, respectively. As the recorded field may therefore not be larger than that of other SNCs, it does not provide clear evidence of an active dynamo at the time of acquisition. However, taking into account the heterogeneous directions observed at the millimeter scale, Weiss *et al*. (2002) have argued that the measured paleointensities underestimate the true paleofield, which could be in the dynamo range. Recall that the age of primary magnetization is likely the largest impact age [4 Ga (Weiss *et al*., 2002)] rather than the magmatic age of 4.5 Ga. The main difficulty in using paleomagnetic data to discuss the age of dynamo shutdown is that the surface field intensity due to crustal remanence can be, locally, as high as the past dynamo field.

PAST LARGE INTERNAL MAGNETIC FIELD: A REQUISITE FOR EARLY MARTIAN LIFE?

There are many possible links between conditions necessary for the appearance of life on Mars and the early large internal magnetic field of Mars. The least speculative is the demonstrated positive effect of an Earth-like magnetosphere on the rate of atmospheric escape due to ion sputtering (see Lammer *et al*., 2003). Various scenarios have been put forward to provide a dense enough atmosphere, an essential condition for life (Carr, 1999). The slower rate of atmospheric escape in the presence of an Earth-like field strongly supports these scenarios. The possibly abrupt transition from a warm to cold dry martian climate has been attributed to the shutdown of the martian dynamo coupled with a decreasing supply of magmatic gases. The age of dynamo shutdown, set quite early in sputtering models (Hutchins *et al*., 1997; Leblanc and Johnson, 2001), may be a key parameter for establishing the length of time life may have reigned on Mars.

The second link is the speculation that living organisms interact with magnetic fields and the early dynamo was a requirement for life. Indeed terrestrial organisms, from bacteria to numerous vertebrates, demonstrate a clear sensitivity to magnetic fields (Wiltschko and Wiltschko, 1995; Kirschvink, 1997). The most apparent manifestation of organisms that interact with magnetic fields is the field-sensitive navigation capacities of a wide range of organisms. This point may be relevant to the evolution of life but not to its appearance, as the very first microorganisms on Earth and perhaps on Mars had probably no need of navigation capacities, being, for example, attached to mineral surfaces.

Another aspect is the shielding from cosmic radiations and associated mutagenic effects by a global magnetic field. This is pertinent to aerial life but not to the most likely early martian life forms, which would have been protected from radiation by their underwater or underground habitat. Moreover, the prediction of increased extinction and appearance of species on Earth during geomagnetic reversals (during which the Earth field is strongly reduced) has never been confirmed despite numerous tests (*e.g.*, Kent, 1977). Lastly, a magnetic field may be a prerequisite for the appearance of life due to its effect on key prebiotic chemical reactions. A breakthrough discovery demonstrated that a magnetic field can produce a chiral imbalance (Rikken and Raupach, 2000). A magnetic field, through its effect on electronic energy levels or by orienting magnetically anisotropic molecules (Weaver *et al*., 2000), may have influenced the complex chemical reactions that led to life. However, the presence of a large field like the present one is not demonstrated at the time of the appearance of life on Earth. Indeed, it has been suggested that strong and stable dynamo action requires the presence of a solid inner core (*e.g.*, Sakuraba and Kono, 1999; Schubert *et al.*, 2000). The inner core of the Earth may have formed only around 2 Ga (Labrosse *et al.*, 1997). The oldest evidence for large paleointensity is 3.5 Ga (Prévot and Perrin, 1992). These two ages suggest that a large (but possibly unstable) field existed at the onset of life on Earth.

APPLICATION OF SURFACE MAGNETIC FIELD MEASUREMENTS

The steady portion of the present field at the surface of Mars is entirely due to the NRM of rocks below the surface down to an estimated depth of 50 km (Langlais *et al.*, 2004). However, this depth is rather arbitrary and actually depends on the magnetic mineral carrying the remanence and the crustal temperature gradient at the time of dynamo shutdown (see discussion in Kletetschka *et al.*, 2000; Rochette *et al.*, 2001, 2005; Dunlop and Arkhani-Ahmed, 2005). Spatial variations in this steady field can be used to delineate the NRM direction and intensity of subsurface formations. This approach of determining NRM by a ground-level survey using a magnetometer has been successfully applied on terrestrial oceanic crust outcrops, on land (Kristjansson, 1993) or by way of a submersible (Macdonald *et al.*, 1983). Depending on the scale of the survey, the "formation" can be a small block or outcrop within the regolith. Indeed, although the loose regolith consists of strongly magnetic materials, bulk NRM intensity will be very low because of the random orientation of the NRM of individual particles. Obviously, the scale of the rover mobility constrains the scale of the NRM that can be surveyed. Mostly, our concept is adapted to blocks in the 5–100-cm scale, protruding from the regolith. For a larger scale (10–1,000 m), an airborne survey is more adapted. To scale NRM and field intensities, one can use the magnetic field (B) generated along its pole by a spherical object of radius r at a distance R: $B = 2/3\ \mu_o\ (r/R)^3$ NRM, with $\mu_o = 4\pi \times 10^{-7}$. At contact with a 1 A/m sphere one thus gets a maximum field of 0.84 μT, and this field will be reduced to 0.1 μT at two radii.

The natural time-varying field is due to atmospheric electric current (in the ionosphere as well as possible electric discharges in the neutral atmosphere) and subsurface electric currents induced by primary atmospheric variations (Menvielle *et al.*, 1996). The measurement of these time variations has direct implications for the human habitability of the martian surface. Indeed, these variations influence the radiation level linked to solar wind charged particles. They can also interfere with communication between a ground base station and an orbiter or the Earth. Moreover, as proposed in the Netlander mission (Menvielle *et al.*, 2000), investigation of induced subsurface currents could allow the identification of a possible solid–liquid water interface and estimation of its depth (thanks to the higher electric conductivity of the liquid phase). This can have major implications for the human habitability of the site and for locating potential survival niches for possible martian life forms.

The proposed MAREDD instrument for MSL included a three-axis fluxgate magnetometer deployed at the end of a boom (Fig. 1) (Rochette *et al.*, 2004a). The magnetometer, designed for various missions, including Mars Netlander (Pedersen *et al.*, 1999; Brauer *et al.*, 2000), has a noise level of 0.01 nT and a range of ±16 μT. These fea-

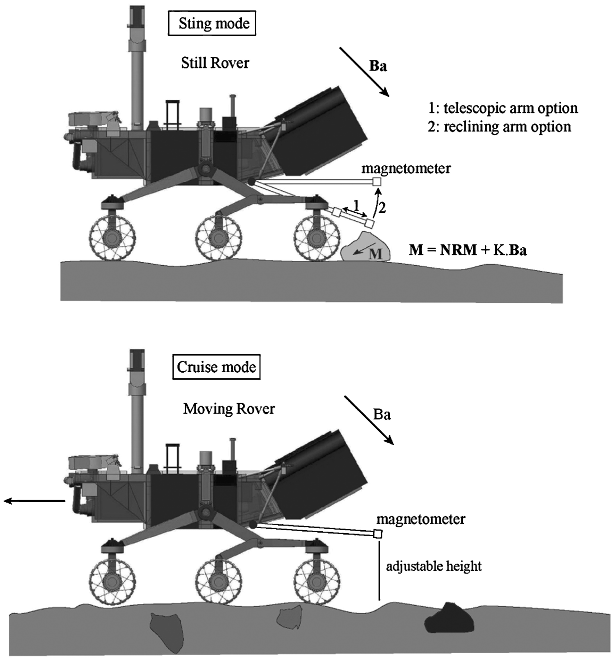

FIG. 1. Sketch of the MSL rover with the proposed placement of the MAREDD instrument at the end of a deployable boom. K, magnetic susceptibility; M, magnetization.

tures may not be suitable for the Mars rover: a higher maximum limit is necessary to avoid saturation over the strongly magnetized Noachian crust, and a lower sensitivity can be chosen because of unavoidable rover noise. The natural field will obviously have to be filtered from the one generated by the rover. This could be accomplished by optimizing the measurement strategy and studying the field generated by the rover and other instruments both before launch and during ground operations on Mars. For the time variation measurements, the instrument could operate in an "observation mode," with every mobile part of the rover perfectly still. This mode of operation may be adopted for short periods during the day and night when no scientific or rover activity takes place. Nevertheless, electric currents due to data transmission or heating devices may create interference and will have to be accounted for.

For spatial variations, two measurement modes can be planned (Fig. 1). During rover displacement, the "cruise mode" will allow the detection of field anomalies linked to strong NRM in the surface or subsurface. The "sting mode" will be used to determine the NRM of specific targets selected from cruise mode results or from the general planning of science activities. The latter mode is much more precise since the rover is still (except for controlled boom deployment and retraction) and contact with the target can be achieved. In this situation, an NRM of 10 A/m generates a field gradient of the order of up to 1 μT. Figure 2 provides an analog of the expected signal using a 20-kg basaltic rock with a natural shape and NRM of 7 A/m. Measurement was performed in the local Earth field (45 μT) using a cesium vapor magnetometer, with the sensor translated within a plane 12 cm above the top of the rock. The contribution of induced magnetization is minor, as we determined a Koenigsberger ratio of 12. For MAREDD, the distinction between spatial and temporal variations could easily be achieved through a "back and forth" measurement sequence, which takes advantage of the mobility of the boom. Spatial variations generated by the rover can be estimated by repeating the operation away from the target. Another option is a fixed gradiometer design (as proposed in Rochette *et al.*, 2004a), optimized to filter the temporal variations but still in need of a strategy to correct for spatial variations.

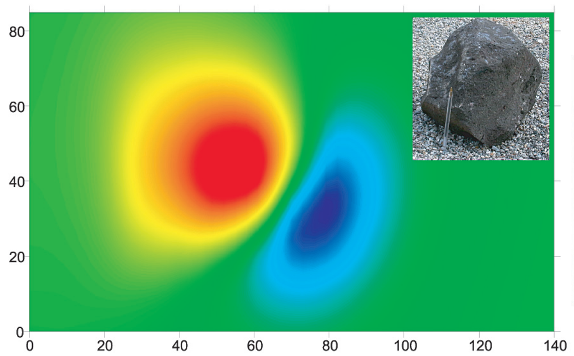

FIG. 2. Magnetic field anomaly map over a magnetized rock (difference between field measured with a moving sensor 12 cm above the rock and a reference sensor placed 2 m away, in nT). The *x* and *y* scales are in cm. **Inset:** A picture of the block.

Two life-related applications of these NRM determinations can be put forward; they are both based on the detection of surface materials with high-intensity NRM.

The first possible interpretation for a strongly magnetized material is that it corresponds to a Noachian age rock. Noachian rock magnetization should be of the order of 10 A/m (Langlais *et al.*, 2004), while more recently magnetized rock will yield an NRM intensity at least one order of magnitude lower. SNC meteorites with established martian magnetization have maximum NRM intensities of 0.4 A/m (Collinson, 1997; Rochette *et al.*, 2001; Gattacceca and Rochette, 2004). To study the events linked to the putative appearance of life on Mars 4 Gyr ago, rocks from this period must be studied and dated individually. Regional surface dating using crater counting will not accurately reflect the age of a given rock found at the surface. Eventually, younger volcanic rocks emplaced on top of rocks characterized by local high-anomaly fields due to Noachian crustal magnetization could reach a high NRM due to the remanent field in the older rock. However, this case should be rare and can be evaluated based on total field measurements.

The second possibility is that the strongly magnetized surface was struck by lightning. Indeed, the large magnetic field pulse associated with lightning current induces a large isothermal remanent magnetization (IRM). According to a survey of martian meteorites, the expected range of saturation IRM is 10–1,000 A/m (Rochette *et al.*, 2005). Lightning NRM is usually a few tenths of

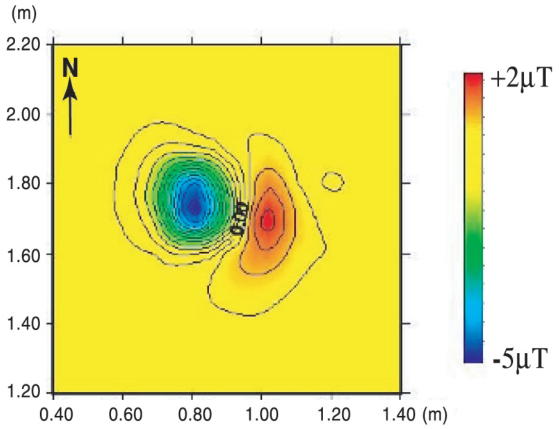

FIG. 3. Magnetic field anomaly (vertical gradient measurement) over a lightning impact on basaltic soil.

saturation IRM (Wasilewsky and Dickinson, 2000; Verrier and Rochette, 2002). The possible occurrence of ground lightning impacts on Mars linked to electric field build up in the atmosphere during dust storms has been put forward by various authors and confirmed by some observations (Farrell and Desh, 2001; Renno *et al.*, 2003). Ground lightning requires a conductive layer at rather shallow depth. The present evidence for widespread evaporitic deposits (Gendrin *et al.*, 2005) and the hypothesis of shallow liquid brines (Burt and Knauth, 2003) fulfill this requirement. A lightning impact is able to remagnetize the surface rock within several square meters, as shown in various terrestrial examples (Fig. 3) (Verrier and Rochette, 2002). As current lines rapidly diverge at depth, remagnetization is confined to the first meter in depth. If lightning impacts are not exceptional on the martian surface, a rover is likely to encounter impacted sites on very old surfaces. For example, the typical lightning impact rate on the Earth is in the range of one to 10 impacts/km^2/year. Assuming a 1 Ma age for the surface, there will be one to 10 impacts per square meter. Although lightning impacts could be a few orders of magnitude less common on Mars, the probability of encountering an impact is heightened because of the much older surface age.

Positive detection of past lightning impacts on Mars has several key atmospheric and biological implications. Lightning impacts are often invoked in the complex reactions that lead to prebiotic molecules. On the other end of the time scale, the probability of lightning is important to assess the present habitability of Mars. Because of the lack of natural trigger points (trees, sharp rocky crests), a human settlement (with antennas, etc.) is likely to attract lightning over large surfaces. Nevertheless, no martian surface probe has been struck by lightning. Note that, if lightning-induced NRM appears widespread on martian surface rocks, the previously mentioned dating objective may not be met. However, one could measure NRMs unaffected by lightning when studying rocks recently excavated by impact cratering. For completeness, alternative sources of lightning on Mars can be invoked: explosive volcanic eruptions and large impacts (Uman, 1987). Their probability of such an encounter by a rover, however, is very limited.

How does one distinguish between lightning-induced and Noachian age NRM? The magnetic field anomaly linked to a homogeneously magnetized Noachian rock is quite simple: it is dipolar if the magnetized rock is an isolated, nearly spherical boulder (Fig. 2). Outcrops struck by lightning, instead, have rapidly varying NRM directions and intensities (Verrier and Rochette, 2002), thereby generating more complex anomalies (Fig. 3) (Jones and Maki, 2005).

A more quantitative discrimination can be obtained using a normalization method. NRM intensity normalized to saturation remanence (IRM) yields the intensity of the magnetic field during remanence acquisition (Verrier and Rochette, 2002; Kletetschka *et al.*, 2003; Gattacceca and Rochette, 2004). A NRM/IRM (REM) ratio of about 10^{-2} is typical of a dynamo, i.e., Earth-like field (10–100 μT range), while smaller fields after the dynamo shutdown (0.1–1 μT range) produces REMs of 10^{-3} and less. In the case of a temporary pulse of a large field during lightning impacts, the REM exceeds 10^{-1}. A normalization method can, therefore, distinguish lightning NRM from Noachian age NRM, and Noachian NRM from younger NRM. Indeed, the actual intensity of NRM depends not only on the intensity of the magnetizing field, but also on the amount and characteristics of magnetic particles in the rock (Dunlop and Ozdemir, 1997). A weak NRM could, therefore, be from a Noachian age rock if this rock is weakly magnetic. This ambiguity is solved through normalization (*i.e.*, the REM technique).

What kind of normalization can be performed in a rover context? One possibility is to use an instrument [like COMASP (see Application of

Surface Magnetic Susceptibility Measurements to the Search for Life)] that provides the magnetic susceptibility (*K*) value for the NRM study target. In simple cases, *K* is proportional to IRM; the NRM/*K* ratio can, therefore, be taken as a proxy for the REM ratio. Another possibility is to have a strong magnet [NdFeB (as in Madsen *et al.*, 1999)] fitted on a contact instrument or sampling arm so that the magnet can be applied to the target. The magnet will impart a local IRM that can be measured with MAREDD. To obtain the equivalent bulk IRM of the target, models of IRM acquisition versus field and of the magnet field geometry are necessary. Although this technique may seem more reliable than the use of COMASP (*i.e.*, the NRM/*K* normalization), the magnet on the rover may disturb the MAREDD instrument.

IRON MINERALS IN MARTIAN REGOLITH

Primary reduced iron-rich minerals on the surface of Mars may have derived from magmatic rocks. According to SNC studies (*e.g.*, Rochette *et al.*, 2001, 2005; Lorand *et al.*, 2005), the major iron-rich phases in these rocks are sulfides (mostly pyrrhotite FeS_{1-x}) and titanomagnetite ($Fe_{3-x}Ti_xO_4$). According to mass balance estimations, up to 10–30% of the regolith mass could be due to meteoritic bombardment (Flynn and McKay, 1990). Meteorites and interplanetary dust particles usually contain 10–20% iron-rich minerals: FeNi metallic alloys, sulfides (mostly troilite FeS), or, more rarely, magnetite. As martian rocks contain only 1–2% iron-rich phases, the major contribution to the superficial stock of primary iron-rich minerals could be meteoritic, *i.e.*, metal or sulfides. Recent analyses by the Mars Exploration Rover (MER) Spirit, which reveal enrichment in Ni (positively correlated with S) in the surface coating of rocks and in soil (Gellert *et al.*, 2004), confirm this hypothesis.

These findings have shed a new light on the possible origin of neoformed Fe^{III} minerals in the regolith, as discussed by Chevrier *et al.* (2004). Since the Viking era (Hargraves *et al.*, 1979), a number of candidates have been proposed for these minerals, including hematite, maghemite, and goethite (*e.g.*, Burns, 1980; Coey *et al.*, 1990; Banin *et al.*, 1993; Morris *et al.*, 1998; Madsen *et al.*, 1999; Christensen *et al.*, 2001; Chevrier *et al.*, 2004). The recent MER results, chiefly based on Mössbauer spectroscopy using MIMOS (Klingelhöfer *et al.*, 1996; Morris *et al.*, 2004), provide new insight. They have identified non-stoichiometric magnetite, a (super)paramagnetic Fe^{III} phase, goethite, and hematite.

The formation of ferric minerals has often been invoked as evidence for the former presence of liquid water and a warm climate, and for the presence of oxidant species (*e.g.*, from oxygen to peroxide and acids). However, 1-year weathering experiments in a $CO_2 + H_2O$ vapor atmosphere (Chevrier *et al.*, 2004) have shown that, while magnetite remains unweathered, metal and sulfide weather very rapidly into goethite ($\alpha FeOOH$), siderite ($FeCO_3$), elemental sulfur, and Fe^{III} sulfate. This rapid weathering of mostly meteoritic minerals in a Mars-like atmosphere, with no liquid water, could be a simple pathway for the production of Fe^{III} phases responsible for the surface color of Mars.

The chemical reactions involved can be summarized for metal as:

$$Fe^0 + 2\,H^+ + CO_3^{2-} \rightarrow Fe^{II}CO_3 + H_2 \quad (1)$$

$$2\,Fe^{II}CO_3 + 2H_2O \rightarrow 2\,Fe^{III}O(OH) + 2\,CO_2 + H_2 \quad (2)$$

and for FeS as (although numerous other species are involved):

$$Fe^{II}S + 2\,H_2O \rightarrow Fe^{III}O(OH) + S^0 + 3/2\,H_2 \quad (3)$$

These spontaneous reactions have been shown to proceed inorganically. However, they can also be mediated by autotrophic anaerobic bacteria. Various goethite-producing autotrophic bacteria or archaea strains grow using FeS (as well as siderite) or metallic iron as electron donors and $CO_2 + H_2O$ (*e.g.*, Dinh *et al.*, 2004; Kappler and Newmann, 2004). According to various authors (Widel *et al.*, 1993; Konhauser *et al.*, 2002), such bacteria are rather archaic and could well have been responsible for the banded iron formations in the early Precambrian, before oxygen appeared because of the development of photosynthesis (see also Kirschvink and Weiss, 2001). Such anaerobic conversion of Fe^{II} into Fe^{III} at a planetary scale could also have occurred on Mars. If so, the recurrent link in literature between water (and indirectly life) and iron mineralogy would be directly substantiated.

Based on this hypothesis and as a guide for future exploration, we suggest where to look for fossil or still viable microorganisms in accessible materials on the martian surface.

This requires the identification of available easily metabolized sources of reduced iron, *i.e.*, FeS and possibly metal. In an already weathered and oxidized surface (*i.e.*, post-Noachian), reduced iron can be found within rocks below the weathering rind. A weathering rind of a few mm (predicted by, *e.g.*, Allen and Conca, 1991) enriched in Fe^{III} was found in the Mazatzal rock investigated by the MER Spirit rover (Gellert *et al.*, 2004; Morris *et al.*, 2004). One could hypothesize that such a weathering rind developed in the past (and possibly up to the present day) through the activity of microorganisms, as shown in terrestrial Mars analog rocks found in hot or cold desert climates (Gorbushina *et al.*, 2002; Onofri *et al.*, 2004). The search for organisms could, therefore, be focused on this weathering rind, especially at the interface with fresh rock. This habitat is indeed favorable: the microporosity of the rock allows access to the atmosphere and to liquid water percolating from melting frost during the warmest time of the day. Biological activity would be extremely slow, as it would proceed during warm periods only, which may be limited to a few days per year [or even per climatic cycle (see Laskar *et al.*, 2004)]. Organisms could, therefore, be sustained even on old (0.1–1 Gyr) surfaces. However, impact gardening periodically regenerates fresh rock surfaces for airborne microorganism colonization. A silicate matrix a few millimeters thick is also a good shelter from ultraviolet and other harmful radiations. Besides minor phases, such as pyrrhotite, there are other more abundant sources of Fe^{II}: pyroxene and olivine. Although direct solubilization of Fe from silicates is known in aerobic bacteria, it is much more demanding (Gillet *et al.*, 2000; Benzezara *et al.*, 2004). However, our pyrrhotite weathering experiment indicated that silicate weathering may be promoted by pyrrhotite reaction products, as occurs in acid mine drainage.

An alternative localized source of reduced Fe, in a much more concentrated form, could result from meteorite in-fall. González-Toril *et al.* (2005) have shown the growth of microorganisms on Earth in an iron meteorite. Bland and Smith (2000) have predicted that an important fraction of the meteorite flux on Mars in the 10–50 g mass range could survive surface impact. The estimated probability of occurrence of a meteoritic fragment (in the 10–50 g range) at the surface is one per 2–200 m^2 (depending on the weathering rate hypothesis). One may wonder why, in the case of such abundance, previous and current surface missions have never detected meteorites, except for the large metallic meteorite discovered by the Spirit rover. As far as we know, small pebbles have never been investigated during previous missions. As it is not possible to drill into or brush a small pebble, the usual targets are either large blocks or loose soil formations. The search for meteorite fragments on the surface of Mars therefore seems feasible in terms of probability, and highly desirable in terms of possible specific niches for microbial survival. In particular, recently fallen meteorites could provide the only accessible source of highly concentrated substrate and may thus be the only target for detectable microorganism contents. Besides autotrophic microorganisms living on sulfide or metal oxidation, carbon-based metabolism could also find a favorable niche in meteorites, which contain not only up to 5% reduced carbon but also amino acids.

APPLICATION OF SURFACE MAGNETIC SUSCEPTIBILITY MEASUREMENTS TO THE SEARCH FOR LIFE

Two possible targets to search for traces of life were proposed in the previous sections. Both targets can be detected or investigated with the proposed COMASP. The COMASP concept derives from the commercial SM30 coil sensor (Rochette *et al.*, 2004a). The calibration and characteristics of measurements using this sensor on natural blocks or surfaces are described in Gattacceca *et al.* (2004). This sensor is based on an oscillator, composed of a 5-cm-diameter coil and capacities that are driven at the LC circuit resonance frequency. The magnetic susceptibility (K) of a material placed in front of the coil is determined by measuring the shift in resonance frequency with respect to the one determined without the sample. Sensitivity of 10^{-7} SI can be achieved, *i.e.*, much higher than the 10^{-3}–10^{-2} SI range of K values in SNCs. Practically 99% of the signal is given by the material present within a 6-cm-thick cylinder 8 cm in diameter placed in front of the coil. In the case of a material with depth-varying

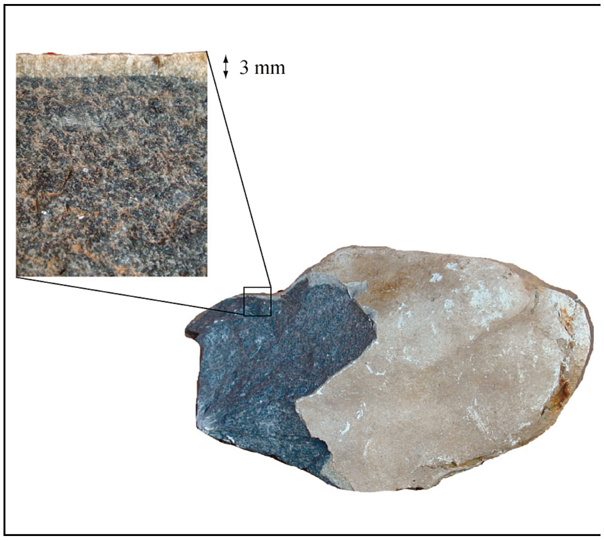

FIG. 4. Weathering rind observed in a large phonolite boulder (from the Vivarais Miocene province), highlighting the white (rind)–dark gray (fresh rock) contrast.

K, the response function favors the more superficial layers. Based on terrestrial analogues [*e.g.*, from Antarctica (Rochette *et al.*, 2004a; Chevrier *et al.*, 2006), the weathering rind is expected to have a large *K* contrast with respect to the unweathered underlying rock. Figure 4 shows another profile studied in a boulder of weathered phonolite (volcanic rock) in a French mountain soil. The susceptibility of the 3-mm-thick rind and fresh rock, determined by subsampling, is, respectively, 112 and 219 $10^{-4} \times$ SI.

The penetration depth and response function of the susceptibility probe can be used to retrieve such a profile of magnetic susceptibility and determine the thickness of the weathering rind and its susceptibility contrast with respect to the underlying bulk rock (or soil) (Gattacceca *et al.*, 2004). This entails measuring *K* with the COMASP coil at different distances from the target. Using measurements at three different distances from the phonolite block, the inversion of SM30 data on the phonolite block (Fig. 4) yields a thickness of 4 mm and susceptibilities of 79 and 186×10^{-4} SI.

The detection of a weathering rind around a rock on Mars before any grinding or coring, and estimation of its thickness and *K* contrast with respect to fresh rock, can help select interfaces that may have been weathered by microorganisms for further, more time-consuming robotic investigation. Indeed, the time required to carry out such determinations using COMASP is less than 1 min (once the instrument touches the target). Once a target is selected using COMASP, a sample of the weathering rind can be obtained by grinding the rock at the depth determined by COMASP.

FIG. 5. Magnetic susceptibility (*K*) measurement using the SM30 on an Antarctic blue ice field during the XX Programma Nazionale di Ricerche in Antartide expedition, showing the different output between a terrestrial schist (left) and a meteorite (right). After Folco *et al.* (2006).

As for meteorite identification on the surface of Mars, our compilation (Rochette *et al*., 2003b, 2004a) shows that the *K* value of most meteoritic materials (chondrites, except R types, and achondrites, except lunar, HED, and angrites, *i.e.*, about 97% of the known meteorite collection) is one or two orders of magnitude larger than that of martian rocks. Systematic measurement of *K* on surface rocks (a few centimeters in size) would allow for the detection of meteorites and an estimation of their concentration in a reasonable amount of time. This concept was validated during an Antarctic meteorite recovery expedition (Fig. 5) (after Folco *et al*., 2006). Suspected meteorites could be further characterized using chemical and mineralogical contact probes, and then processed as a whole (because of their small size) in an analytical laboratory, with particular attention to possible biological material.

DISCUSSION AND CONCLUSIONS

The review part of the present contribution does not include an update on the hypothesis of fossil traces of biological activity in ALH84001 (McKay *et al*., 1996), based in particular on the interpretation of its magnetite crystals as fossils of magnetotactic bacteria (Thomas-Keprta *et al*., 2000). Indeed, detailed discussions can be found in various recent papers, most of them concluding an inorganic origin (Barber and Scott, 2003; Brearley, 2003; Treiman, 2003; Golden *et al*., 2004), whereas McKay *et al*. (2003) have brought renewed support for the biogenic interpretation. Various papers related specifically to magnetism of ALH84001 have been published (Kirschvink *et al*., 1997; Weiss *et al*., 2000, 2002, 2004; Antretter *et al*., 2003; Rochette *et al*., 2005). We have reviewed the wealth of data and ideas related to magnetic fields, magnetic properties, and iron minerals that have stemmed recently from spatial exploration and from the laboratory study of martian meteorites and analogs of martian materials and surface processes. We have also shown that these data are somewhat relevant to the broad topic of life on Mars.

An Earth-like magnetic field generated by a dynamo was active during the first 0.3–0.7 Gyr of martian history. This field may have played a key role in preventing the escape of the dense atmosphere, thus ensuring the longevity of a warm and wet climate favorable to the appearance of life. Direct links between possible early life and the magnetic field (through magnetotactic phenomena and the influence of the magnetic field on prebiotic or biological chemical reactions) may be invoked but are much more speculative.

Magnetic field measurements in future missions to the surface of Mars would have various applications of astrobiological interest. They would allow: (1) identification of strongly magnetized rocks that are either of Noachian age or have been struck by lightning and (2) detection of the subsurface solid–liquid water interface through electromagnetic sounding.

A number of iron-bearing minerals have been detected or hypothesized to exist in the martian regolith, but their origin is still debated. Besides invoking a very ancient warm and wet climate for all oxidized ferric phases, formed at the expense of martian rocks, we suggest an alternative (but not exclusive) hypothesis that iron weathering is recent and purely atmospheric, involving the alteration of metal of meteoritic origin and sulfide of either internal or meteoritic origin. In this case, secondary minerals like hematite, goethite, jarosite, etc., cannot be invoked as evidence of abundant liquid water. However, the oxidation of metal and sulfide in a $CO_2 + H_2O$ atmosphere is efficiently mediated by known terrestrial bacteria. In the hypothesis that similar bacteria appeared on Mars, there are two possible targets for finding martian life forms: (1) recently fallen meteorites on the martian surface and (2) the interface between fresh sulfide-bearing rocks and their weathering rinds. Both targets can be detected and characterized using a contact magnetic susceptibility probe.

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ABBREVIATIONS

COMASP, contact magnetic susceptibility probe; IRM, isothermal remanent magnetization;

MAREDD, magnetic remanence device; MER, Mars Exploration Rover; MSL, Mars Science Laboratory; NRM, natural remanent magnetization; REM, natural remanent magnetization/isothermal remanent magnetization.

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