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Future Mars geophysical observatories for understanding its internal structure, rotation, and evolution

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A B S T R A C T

Our fundamental understanding of the interior of the Earth comes from seismology, geodesy, geochemistry, geomagnetism, geothermal studies, and petrology. For the Earth, measurements in those disciplines of geophysics have revealed the basic internal layering of the Earth, its dynamical regime, its thermal structure, its gross compositional stratification, as well as significant lateral variations in these quantities. Planetary interiors not only record evidence of conditions of planetary accretion and differentiation, they exert significant control on surface environments.

We present recent advances in possible in-situ investigations of the interior of Mars, experiments and strategies that can provide unique and critical information about the fundamental processes of terrestrial planet formation and evolution. Such investigations applied on Mars have been ranked as a high priority in virtually every set of European, US and international high-level planetary science recommendations for the past 30 years. New seismological methods and approaches based on the cross-correlation of seismic noise by two seismic stations/landers on the surface of Mars and on joint seismic/orbiter detection of meteorite impacts, as well as the improvement of the performance of Very Broad-Band (VBB) seismometers have made it possible to secure a rich scientific return with only two simultaneously recording stations. In parallel, use of interferometric methods based on two Earth–Mars radio links simultaneously from landers tracked from Earth has increased the precision of radio science experiments by one order of magnitude. Magnetometer and heat flow measurements will complement seismic and geodetic data in order to obtain the best information on the interior of Mars.

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In addition to studying the present structure and dynamics of Mars, these measurements will provide important constraints for the astrobiology of Mars by helping to understand why Mars failed to sustain a magnetic field, by helping to understand the planet's climate evolution, and by providing a limit for the energy available to the chemoautotrophic biosphere through a measurement of the surface heat flow. The landers of the mission will also provide meteorological stations to monitor the climate and obtain new measurements in the atmospheric boundary layer.

1. Introduction

1.1. Why study Mars?

By studying other planets, we seek to understand the processes that govern planetary evolution and discover the factors that have led to the unique evolution of Earth. Why is Earth the only planet with surface liquid oceans, plate tectonics, and abundant life? Mars is presently on the edge of the habitable zone, but may have been much more hospitable early in its history. Recent surveys of Mars suggest that the formation of rocks in the presence of abundant water was largely confined to the earliest geologic epoch, the Noachian age (prior to 3.8 Ga) (Poulet et al., 2005). This early period of Martian history was extremely dynamic, witnessing planetary differentiation, formation of the core, an active dynamo, the formation of the bulk of the crust and the establishment of the major geologic divisions (Solomon et al., 2005). Formation of the crust and associated volcanism released volatiles from the interior into the atmosphere (Greeley, 1987; Phillips et al., 2001; Gillmann et al., 2009), causing conditions responsible for the formation of the familiar signs of liquid water on the surface of Mars, from the abundant channels, phyllosilicate formations and carbonate deposits to sulfate-rich layered outcrops (Poulet et al., 2005; Clark et al., 2005; Bridges et al., 2001; Ehlmann et al., 2008; Morris et al., 2010).

1.2. Why study the geophysics of Mars?

Studying the geophysics of Mars, focusing on interior processes and early evolution, provides essential constraints for models of the thermal, geochemical, and geologic evolution of Mars and for the correct use of the constraints from SNC meteorites (considered to be Martian rocks, based on a close match between the composition of gases included and the atmosphere of Mars, often used in the literature to constrain the Martian mantle composition; e.g., Wänke and Dreibus, 1988; Hauck and Phillips, 2002) and any future samples from Mars.

Our fundamental understanding of the interior of the Earth (and of the Moon) comes from geophysics, geodesy, geochemistry, and petrology. For geophysics, seismology, geodesy, and heat flow measurements have revealed the basic internal layering of the Earth, its thermal structure, its gross compositional stratification, as well as lateral variations in these quantities. For example, seismological observations effectively constrained both the shallow and deep structure of the Earth at the beginning of the 20th century, when seismic data enabled the discovery of the crust–mantle interface (Mohorovičić, 1910) and measurement of the outer core radius (Oldham, 1906), with a 10 km accuracy (Gutenberg, 1913). Soon afterwards, tidal measurements revealed the liquid state of the outer core (Jeffreys, 1926), and the inner core was seismically detected in 1936 (Lehmann, 1936). A similar sequence has been followed for the Moon with the Apollo seismic and geodetic data, with the crust determination from artificial and natural impacts (e.g., Toksöz, 1974; Toksöz et al., 1972; Chenet et al., 2006) to mantle structure (Nakamura, 1983; Khan and Mosegaard, 2002; Lognonné et al., 2003; Gagnepain-Beyneix et al., 2006) and more recently to the determination of the core size and state with tidal measurement (Williams et al., 1996, 2001) and modern seismic processing (Weber et al., 2011; Garcia et al., 2011). Subsequently, with the development of seismic networks on the Earth, seismology has mapped the structure of core–mantle boundary, density and phase changes in the mantle, three-dimensional velocity anomalies in the mantle related to sub-solidus convection, and lateral variations in lithospheric structure. Additionally, seismic information places constraints on Earth's interior temperature distribution and on the boundary conditions at the top and bottom of the outer core, which govern the mechanisms of geodynamo operation (e.g., Aubert et al., 2008). Thus physical properties inferred from seismic data are used in almost any modeling of the Earth's thermal and volatile evolution, including the exchange of volatiles among different reservoirs (McGovern and Schubert, 1989; van Keken and Ballentine, 1999; Franck and Bounama, 1997, 2000; Schubert et al., 2001; Guest and Smrekar, 2007; Smrekar and Guest, 2007), and their impact on the long term habitability of the planet.

The main difference between the Earth and Mars is that the latter still preserves many billion year old crustal and potentially mantle structures on a planetary spatial scale (e.g., the heavily cratered southern hemisphere of Mars), while the ocean floor that covers about two-thirds of the Earth's surface is younger than 250 million years, due to plate tectonics and associated recycling of the Earth's lithosphere. Martian meteorite compositions indicate melting source regions with different compositions that have persisted since the earliest evolution of the planet (Jones, 1986; Borg et al., 1997, 2002). Further, much of the Martian crust dates to the first half billion years of solar system history (Frey et al., 2002; Frey, 2006a, 2006b). Measurements of the interior are likely to detect mantle inhomogeneities that still reflect differentiation and early planetary formation processes, making Mars an ideal subject for geophysical investigations aimed at understanding planetary accretion and early evolution.

Subsequent to initial differentiation, Mars and the Earth diverged in their evolution. Earth's thermal engine has transferred heat to the surface largely by lithospheric recycling over much of its history, but on Mars there is no evidence in the available record that this process ever occurred (e.g., Pruis and Tanaka, 1995; Sleep and Tanaka, 1995). Over the past $\sim$4 billion years, giant hotspots (Tharsis and Elysium) have played a significant role in the tectonic, thermal and volatile evolution of the planet, and are possibly related to an early core dynamo (Johnson and Phillips, 2005), which may, in turn, have been crucial for shielding Mars' early atmosphere from solar wind erosion (Kallio and Janhunen, 2001, 2002; Fang et al., 2010a, 2010b; Lammer et al., 2008, 2009). Furthermore, these volcanic complexes released massive amounts of volatiles to the Martian atmosphere, which possibly led to clement conditions at times and provided temporary habitable environments (Phillips et al., 2001).

Geophysical studies will provide important constraints for the astrobiology of Mars by helping to understand why Mars failed to maintain a global magnetic field, why Mars has undergone such dramatic changes in climate over its history, and by providing a limit for the energy available to the chemoautotrophic biosphere through a measurement of the heat flow.

Planetary interiors not only record evidence of conditions of planetary accretion and differentiation, they exert significant control on surface environments. The structure of a planetary interior and its dynamics control heat transfer within a planet through advected mantle material, heat conducted through the lithosphere, and volcanism. Volcanism in particular controls the timing of volatile release, and influences the availability of water and carbon.

Crust: The crust of a planet is generally thought to form initially through fractionation of an early magma ocean, with later addition through partial melting of the mantle and resulting volcanism. Thus the volume (thickness) and structure (layering and lateral variations) of the crust, along with the composition of the mantle, places strong constraints on the depth and evolution of the putative Martian magma ocean and, by extension, planetary magma oceans in general (Elkins-Tanton et al., 2003, 2005a). Currently we do not know the volume of Mars' crust to within a factor of two (Wieczorek and Zuber, 2004). Orbital data allows the calculation of relative variations in crustal thickness (Neumann et al., 2004), but these models require many assumptions, in particular a tie point to allow conversion of these relative variations to absolute values (such as an assumed mean thickness) and densities of the crust and mantle.

Mantle: Mantle dynamics plays a key role in shaping the geology of the surface through volcanism and tectonics (Van Thienen et al., 2007). The radius of Mars' core has implications for possible mantle convection scenarios and in particular for the presence of a perovskite phase transition at the bottom of the mantle, which enables global plume-like features to exist and persist over time (Harder and Christensen, 1995; Breuer and Spohn, 2006; Spohn et al., 1998). Such strong, long-standing mantle plumes arising from the core–mantle boundary may explain the long-term volcanic activity in the Tharsis area, but their existence during the last billion years is uncertain. An alternative for Tharsis volcanism is that thermal insulation by locally thickened crust, which has a lower thermal conductivity and is enriched in radioactive elements, leads to a significant temperature increase in the upper mantle sufficient to generate partial melt (Schumacher and Breuer, 2006). We note that the tidal Q of Mars is $\sim$80 (Smith and Born, 1976; Bills et al., 2005; Lainey et al., 2007; Jacobson, 2010), substantially less than that of the Earth's mantle ($\sim$200–280), meaning that Mars is surprisingly dissipative.

Core: Knowledge of the core state and size is crucial for understanding a planet's history, and the thermal evolution of a terrestrial planet is determined by the dynamics of its mantle and core. The possibility of magnetic field generation in a liquid core is dependent on the planet's ability to develop convection in the core that has the appropriate motions to generate magnetic fields and that is sufficiently vigorous to overcome ohmic dissipation. In particular, a core dynamo depends on the heat flow at the core–mantle boundary, a high thermal gradient in the liquid core, or latent heat and/or light elements released during the growth of a solid inner core (Longhi et al., 1992; Dehant et al., 2007; 2009; Breuer et al., 2007). The state of the core depends on the percentage of light elements in the core and on the core temperature, which is related to the heat transport in the mantle (Stevenson, 2001; Breuer and Spohn, 2003, 2006; Schumacher and Breuer, 2006; Dehant et al., 2009). Thus the present size (and state) of the core has important implications for our understanding of the evolution and current state of Mars (Breuer et al., 1997; Stevenson, 2001; Spohn et al., 2001a; Van Thienen et al., 2006, 2007; Dehant et al., 2007, 2009, 2011). However, the value of the core radius is currently uncertain to at least $\pm 15\%$ (Rivoldini et al., 2011) and it is unclear whether it is liquid or if it hosts a solid inner core (Yoder et al., 2003). While the indications from Phobos Q (Lognonné and Mosser, 1993; Zharkov and Gudkova, 1997) and measurements of Love number from recent Mars satellite geodesy (Konopliv et al., 2006, 2011; Marty et al., 2009) are consistent with a core that is at least partially liquid, direct detection from seismic reflected phases is necessary to confirm this result. Based on the expected mantle attenuation, the low Q determined from the secular evolution of the orbit of Phobos (Bills et al., 2007) strongly supports a liquid core as well (Lognonné and Mosser, 1993). Core structure, in particular its size and the possible existence of an inner core, plays a central role in determining the history and strength of any planetary magnetic field (Mocquet et al., 2011).

1.3. *What kind of instruments can help geophysical studies?*

The payload required to provide answers to these questions has already been identified in several previous studies in Europe and USA focused on network missions to Mars. The instruments identified to meet these geophysical objectives are the following (the short names of the experiments used in this paper are given in parentheses):

- Seismometer (SEIS),
- Heat flow probe (HP^3, for Heat Flow and Physical Properties Package),
- Radio-science Geodesy Experiment (RISE, for Rotation and Interior Structure Experiment),
- Magnetometer (MAG).

The suite of instruments listed above should be complemented by an atmospheric sensor Package (ATM) and a stereo panoramic imaging system (SPCAM, for Stereo Panoramic Camera), which are necessary to support the deployment of the geophysical payload and its complete analysis, including geological context of the landing site and meteorological decorrelation of seismic data (e.g., Beauduin et al., 1996). Additional science objectives of these two instruments are not addressed in this paper. As the geophysical instruments require long-term measurements, and as atmospheric studies benefit from measurements over at least a full seasonal cycle, a mission operational lifetime of at least one Martian year (687 days) is needed.

Mars Geophysical Network projects have received repeated recommendation by the International Mars Exploration Working Group (IMEWG) and have been fully endorsed by the worldwide scientific community on many occasions. These landers would probe the interior of the planet through seismic monitoring, magnetic sounding, heat flow measurements, and measurements of its rotational dynamics. They would study the global circulation and regional dynamics of the atmosphere and through imaging would reveal the geology of a host of new places on the surface. The enthusiasm for this endeavor has been manifested in a number of mission studies including Mars Environmental Survey (MESUR) (Solomon et al., 1991; Hubbard et al., 1993; Squyres, 1995), MarsNet (Chicarro et al., 1991; 1993), and InterMarsnet (Chicarro et al., 1994; Banerdt et al., 1996). At the beginning of the 21st century the goal was close to being realized with the NetLander mission (Lognonné et al., 1998; Harri et al., 1999; Counil et al., 2001; Marsal et al., 2002; Dehant et al., 2004), which was sponsored by a broad consortium of nations in Europe and America. Unfortunately, budget difficulties and programmatic conflicts at CNES and NASA resulted in its cancellation. A new attempt was proposed for the ExoMars mission platform, the GEP, the Geophysical Package (Lognonné, 2005; Ulamec et al., 2007). However, for resource allocation reasons, the GEP has not been selected to be part of the future ExoMars dual-mission. Thus 15 years after the failure of Mars 96, no recovery of that mission geophysical science or geophysical network mission of any sort has been implemented for flight (Linkin et al., 1998).

However, the scientific priorities for understanding the origin, evolution, environment and habitat of Mars remain and geophysical landers are clearly required to make the measurements necessary to address these questions. A large group of scientists in Europe, North America and Asia remain committed to goal of deploying an International Geoscience Observatory on Mars, whose data would be available to the world science community. For example, in order to propel forward the geophysical investigation of Mars and provide constraints on internal structure and tectonic activity, US scientists, with partners in Europe, Canada and Japan, have recently proposed a Discovery mission (the GEophysical Monitoring Station (GEMS) mission), comprising a geophysical station that would serve as a one-lander network precursor (Banerdt et al., 2010). This mission was selected in May of 2011 (along with two other missions) for a Phase A study, and will be subject to a competitive down-selection for final approval in mid-2012.

A promising development is that NASA and ESA are co-aligned in their interests for the exploration of Mars over the next decade (i.e., a Trace Gas Orbiter Mission, landed rovers, and network mission concepts) and beyond (sample return). The US Decadal Survey (2011) recommended recently a Mars program "taking the first critical steps toward returning carefully selected samples from the surface of Mars", and endorsed the value of Mars interior science. Within the ESA possible next missions, there is the possibility to have, in 2020–2022 timeframe, a geophysical mission performing, for the first time, an in-situ investigation of the interior of a truly Earth-like planet other than our own. Such a mission provides unique and critical information about the fundamental processes of terrestrial planet formation and evolution. This investigation has been ranked as a high priority in virtually every set of US and international high-level planetary science recommendations for the past 30 years (e.g., COMPLEX (COMmittee on Planetary and Lunar EXploration), 1978, 1994, 2003; Space Studies Board, 1988, 2006; Decadal Survey, 2003; NOSSE (New Opportunities in Solar System Exploration: An Evaluation of the New Frontiers Announcement of Opportunity), 2008; MEPAG (Mars Exploration Program Analysis Group), 2008). In particular, such a mission would begin the geophysical exploration of the Martian interior using seismic and thermal measurements and rotational dynamics, providing information about the initial accretion of the planet, the formation and differentiation of its core and crust, and the subsequent evolution of the interior.

As has been documented in many places (e.g., NetSAG (Mars Network Science Analysis Group), 2010), multiple landers making simultaneous measurements (a network) are required to fully address the objectives for understanding terrestrial planet interiors. However, a pair of (or more) geophysical stations is still valuable as measurements constraining the structure and processes of the deep interior of another planet are virtually nonexistent. In addition, these two stations might be supplemented by other additional stations in subsequent missions from Europe, US, Japan, or China, such as the GEMS mission above. Even with only two landers, a geophysical mission provides groundbreaking measurements resulting in a significant leap in our understanding of a wide range of previously unexplored areas, as it will be demonstrated in this paper. Science performed with a single geophysical station will also provide key constraints on the crustal and subsurface context (e.g., heat flow, porosity profile, water profile, crustal thickness). The application of sophisticated, state-of-the-art, single-station analysis techniques to high-quality broad-band seismic data for instance allows the extraction of planetary parameters such as crustal thickness, core state, and seismic activity level (results that can be derived from "guaranteed" non-traditional signals such as meteorite impacts, Phobos tide, and atmospheric interaction with the surface). A seismic network would allow studying the geographic distribution of seismicity and identifying active areas, source processes, spatial variations of the interior structure, and would generally increase resolution and reduce ambiguities.

Future Sample Return mission from a given location will benefit enormously from the prior deployment of a geophysical station at that site providing a full local contextual understanding (geophysical, geological, hydrological, surface, atmospheric) for the returned samples.

2. Science context

2.1. Main science questions

The science goals of geophysics of Mars are stated as follows:

1. Understand the formation and evolution of terrestrial planets through investigation of interior structure and processes of Mars.
2. Determine the present level of tectonic activity and impact flux on Mars.

From these high level goals, a fundamental set of baseline science objectives (Table 1) can be derived.

To accomplish these objectives, a tightly focused payload has been identified consisting of several payload elements. These instruments are a seismometer (e.g., the Seismic Experiment for Interior Structure—SEIS, Lognonné et al., 2000), a magnetometer (e.g., the MAGnetic measurements instrument—MAG, Menvielle et al., 2000), a heat flow probe (e.g., the Heat Flow and Physical Properties Package—HP3, Spohn et al., 2001b), and a precision radio tracking geodesy experiment (e.g., the Rotation and Interior Structure Experiment—RISE, Dehant et al., 2009, 2011). The landers will also have atmospheric sensors to measure the meteorological parameters such as temperature, pressure and wind, providing information necessary for interpreting the seismic observations as well as providing constraints for general circulation models (GCMs). Additional payload elements may be included to support these investigations, such as a deployment arm and a camera.

The above-listed goals and objectives flow directly from numerous planetary science planning documents, as for example, the US Decadal Survey (2003). The International Academy of Astronautics (IAA) Planetary Study Group Report lists the expansion of the human horizon in space (human exploration, Astronauts to Mars) and the search for extraterrestrial life as primary goals of Exploration. This report lists Sample Return and Network Missions as mandatory. The IAA 'Heads of Space Agencies' summit declaration of November 2010 has acknowledged this report as part of a high priority effort to work together to achieve the next leap in understanding of our Solar System.

Table 1
List of science objectives.

- Determine the size, composition and physical state of the core.
- Determine the mineralogy, porosity and thickness of the crust from geophysical parameters (seismic velocities, thermal and electrical conductivity).
- Determine the mineralogy of the mantle from geophysical parameters (seismic velocities, density, thermal and electrical conductivity).
- Determine the thermal state of the interior and constrain the distribution of radiogenic elements with depth.
- Measure the rate and distribution of internal seismic activity.
- Measure the rate of impacts on the surface.
- Constrain the knowledge of a possible inner core.

2.2. State of the art in the research field

2.2.1. Early history of Mars

The thermal and chemical evolution of Mars during its earliest history, when it was most geologically active, is poorly constrained. Mars displays a striking hemispherical dichotomy in both topography and crustal thickness that separates the heavily cratered southern highlands from the smooth northern lowlands. There are numerous hypotheses for how the dichotomy formed. Formation by an oblique giant impact is currently favored (Andrews-Hanna et al., 2008), although endogenic models such as plate tectonics (Sleep, 1994; Breuer and Spohn, 2003; Lenardic et al., 2004) and degree one mantle convection (Wise et al., 1979a, 1979b; Zhong and Zuber, 2001) cannot be ruled out with current information. Mantle overturn is also thought to lead to substantial re-melting in the deepest mantle, which may have influenced early Martian processes such as the development of the crustal dichotomy (Debaille et al., 2009). The existence of crustal remnant magnetization on Mars (Connerney et al., 1999) indicates that a dynamo operated for a substantial time early in Martian history, but the timing, duration, and driving mechanism are unknown. Hypotheses include inner core formation (Stevenson, 2001), an early hot core (Breuer and Spohn, 2003; Williams and Nimmo, 2004) and plate tectonics (Breuer and Spohn, 2003; Nimmo and Stevenson, 2001).

The three leading models for the thermal evolution of Mars are stagnant lid convection (Hauck and Phillips, 2002), early plate tectonics followed by stagnant lid convection (Breuer and Spohn, 2003; Lenardic et al., 2004; Nimmo and Stevenson, 2001; Spohn et al., 2001a), and global mantle overturn (Elkins-Tanton et al., 2003, 2005a; Parmentier and Zuber, 2007; Zaranek and Parmentier, 2004). Each model has different predictions for the evolution of interior heat flux with time and its current state (Fig. 1), the period of dynamo activity (Nimmo and Stevenson, 2001; Hauck and Phillips, 2002; Elkins-Tanton et al., 2005b), and the timing of formation and thickness of the crust (Hauck and Phillips, 2002). Unraveling the story of crustal formation and early convection requires investigating the major aspects of crustal structure that reflect the processes that formed it: crustal thickness, large-scale layering, and the global topographic dichotomy, the feature that, along with the Tharsis rise, dominates Martian tectonic and convective history.

2.2.2. Crustal thickness

Crustal thickness is a sensitive indicator of the thermal and dynamic evolution of a planet. For example, plate tectonics, stagnant lid, and mantle overturn models predict thin, medium, and thick crust, respectively. Predictions for crustal thickness from parameterized convection models (Hauck and Phillips, 2002; Morschhauser et al., 2011) range from 10 s to 100 s of km for plausible variables. Nominal models give thicknesses of $\sim$60 km, which is $\sim$75% emplaced at 4.0 Ga. Predicted thicknesses are a very strong function of the water content of the mantle and the initial mantle temperature. Although plate tectonics is not explicitly included in the models by Hauck and Phillips (2002), a range of mantle temperatures that encompasses the reduced mantle temperature predicted by plate tectonics is considered. The effect is to significantly reduce the amount of crust produced. Models of crustal formation via global overturn are most consistent with average crustal thicknesses of $\sim$100 km (Elkins-Tanton et al., 2005a).

Gravity and topography data provide some constraints on the Martian crustal thickness. But without seismic data, there is a factor of 2–3 uncertainty in the estimate due to differences in assumptions and methods. Using the Bouguer gravity and assuming a non-zero crustal thickness at Hellas gives an average crustal thickness of $>$45 km (Neumann et al., 2004), whereas gravity/topography ratio analysis suggests 57 ± 24 km, localized admittance suggests a value around 50 km, and topographic relaxation gives an upper bound of 115 km (see Wieczorek and Zuber, 2004). Estimates of moment of inertia (MOI) and core density starting from a plausible mantle mineralogy and an Fe–FeS liquid core have been used to place an upper limit of 260 km on the crustal thickness (Kavner et al., 2001; Rivoldini et al., 2011).

2.2.3. Core size and composition

The size of the core has major consequences for internal structure and planetary evolution. For example, for a cold mantle (temperatures below 2000 K) a large core makes a perovskite-bearing lower mantle impossible, due to insufficient pressure at the base of the mantle. The endothermic phase transition spinel-to-perovskite has a strong effect on mantle convection. The presence of such a phase change could explain the formation of Tharsis during Mars' early history from a single-plume convection pattern (Harder and Christensen, 1996; Van Thienen et al., 2005). The size and composition of the core are also important in the history of the magnetic dynamo, which in turn has important consequences for the retention of the atmosphere and the possible habitability of the surface early in Mars' history.

When the temperature (cold or warm mantle) and the mineralogy of the mantle is fixed, the moment of inertia is a strong constraint on the core size, density and temperature, and mantle mineralogy, with weaker constraints on the crustal thickness and the sulfur concentration in the core (Sohl et al., 2005). An example of the range of possible core sizes for a range of plausible core and mantle compositions is shown in Fig. 2.

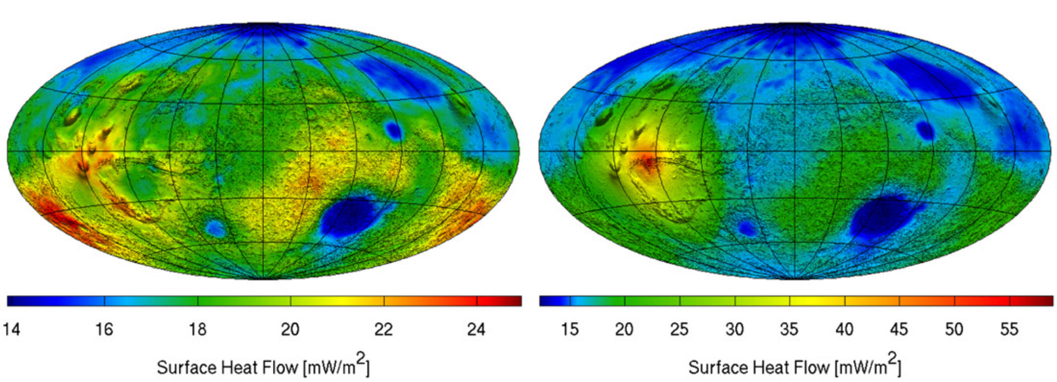

Fig. 1. Left: Spatial variability of the surface heat flux expected from variations of the local crustal thickness and the local abundance of heat producing elements. Right: Spatial variability of the surface heat flux if a stable mantle plume is active underneath Tharsis. Figure similar to Fig. 6 in Grott and Breuer (2010), but using the nominal thermal evolution model of Morschhauser et al. (2011).

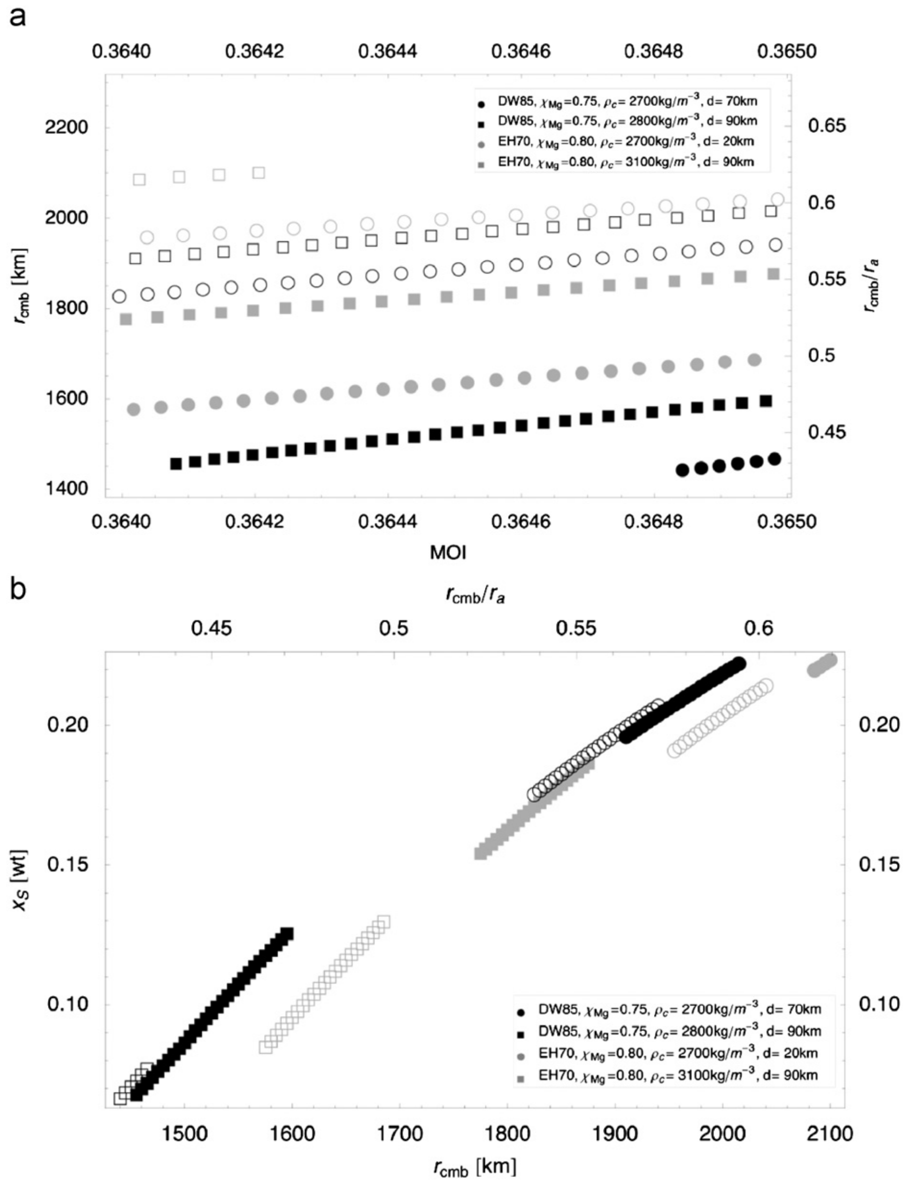

Fig. 2. (a) Core radius r_{cmb} vs. normalized average moment of inertia (MOI) and (b) sulfur ratio X_S vs. core radius r_{cmb} for different mantle compositions and end-member cold and hot mantle temperatures. Open circles and boxes are for hot mantle temperature end-member, filled boxes and circles are for cold mantle temperature end-member. Mantle mineralogies are those of Dreibus and Wänke (1985) (DW85) and of Sanloup et al. (1999) (EH70). For each of the two mineralogical models circles are for smallest MOI-compatible crust density ρ_c and thinnest crust thickness d, boxes are for largest and thickest. r_{cmb} is chosen such that core sulfur weight fraction is $X_S \leq 22\%$. χ_{Mg} is the magnesium number of the mantle.

The only "direct" observations of the interior are those of the gravity field and polar MOI from spacecraft radio tracking (Smith et al., 1993; Folkner et al., 1997; Konopliv et al., 2006, 2011; Marty et al., 2009). These observations are the main existing constraints for interior models. Without additional constraints for mantle mineralogy (e.g., from SNCs) or for the core composition (e.g., the percentage of light elements in the core), the core radius estimate ranges from 1300 to 2200 km. The relative enrichments of iron in the Martian mantle and sulfur in its core with respect to the Earth's depend strongly on the initial composition and differentiation history of the planet. Theoretical calculations of planetary thermal evolution that incorporate these models and fulfill the measured polar MOI lead to outer core radius values ranging from 1600 to 1850 km (Schubert and Spohn, 1990; Schubert et al., 1990, 1993; Mocquet et al., 1996; Sohl and Spohn, 1997; Folkner et al., 1997; Zharkov and Gudkova, 2000; Verhoeven et al., 2005; Mocquet et al., 2011; Rivoldini et al., 2011). Additional constraints from orbital observation of the tidal Love number k_2 reduce further the range of values. The latest k_2 value and its uncertainties (Konopliv et al., 2011) provide a range (at 1σ) of core radii between 1750 and 1950 km (Rivoldini et al., 2011) (see Fig. 3).

Analyses of the measured tidal effect on spacecraft orbits suggests that at least the outer portion of the Martian core is liquid (Yoder et al., 2003; Balmino et al., 2006; Konopliv et al., 2006, 2011; Marty et al., 2009). However its interpretation in terms of the existence or not of an inner core is on the edge of the present capabilities from orbit, and requires independent confirmation using surface-based measurements.

2.2.4. Magnetic field

There is currently no global planetary magnetic field at Mars of intrinsic origin, but the discovery of intense local magnetic anomalies in the heavily cratered southern hemisphere terrain

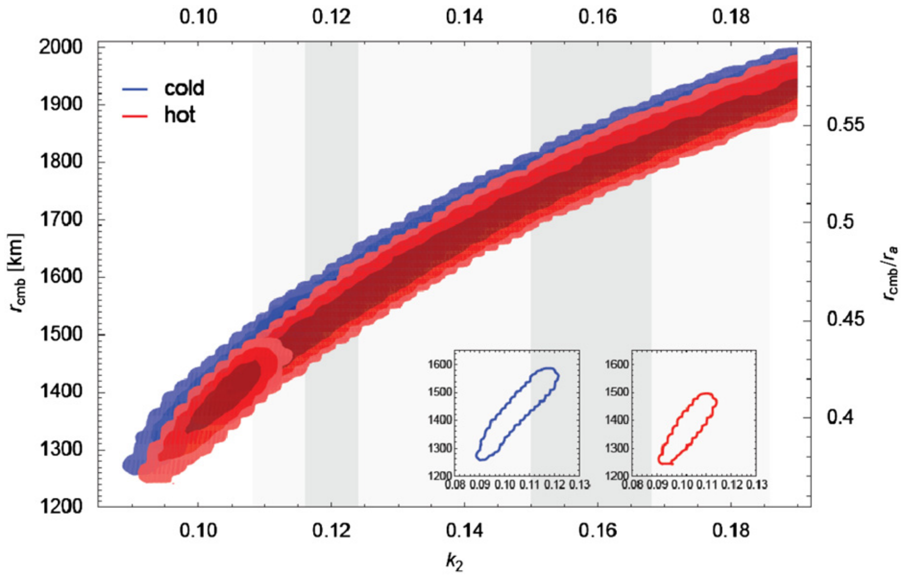

Fig. 3. The tidal Love number k_2, when combined with MOI measurements, provides strong constraints on core radius. Plot shows core size versus k_2 for all possible mantle and core mineralogies and for both hot (red) and cold (blue) mantles (similar to Rivoldini et al., 2011); recent k_2 ranges (Marty et al., 2009; Konopliv et al., 2011) are shown in gray (values of k_2 with 1σ) and in light-gray (values of k_2 with 3σ). The two insets show as well the results for models with an inner core. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

by the Mars Global Surveyor (MGS) Magnetometer investigation revealed evidence for magnetized crust, and by inference an ancient global magnetic field (i.e., a past core dynamo) (Acuña et al., 1998, 1999, 2001; Cloutier et al., 1999; Brain et al., 2003; Vennerstrøm et al., 2003; Johnson and Phillips, 2005; Quesnel et al., 2007; Kletetschka et al., 2009). With in situ magnetic field measurements we cannot distinguish the details of the mechanisms by which the rocks were magnetized (e.g., thermal versus chemical versus detrital magnetizations), but we could set constraints on the volume and depth of the magnetized material and the magnetization intensity.

The magnetic field is a sum of contributions from various sources, most notably the crustal remanent magnetization, electric currents in the plasma environment, and electric currents induced in the Martian crust and upper mantle. To understand the signal from internally induced currents, we can use electromagnetic sounding. The depth of penetration of an electromagnetic wave in a conductive medium depends on both the period of the studied phenomenon and the electrical resistivity of the medium: the longer the period and the higher the resistivity, the greater the depth of penetration (skin depth). Thus higher frequencies spectrum enable one to probe the uppermost kilometers of the crust, and lower frequencies enable one to probe the mantle down to a few hundred kilometers (Menvielle et al., 1996, 2000; Mocquet and Menvielle, 2000; Grimm, 2002; Langlais and Quesnel, 2008; Langlais et al., 2004, 2009, 2010).

The electrical resistivity of geological materials varies greatly with temperature and the percentage of conductive fluids (molten rocks, water rich fluids) within the solid matrix. For non-hydrated rocks, the resistivity remains very high for temperatures up to 1200 °C or even 1800 °C in some cases. It is generally on the order of, or greater than a few tens of thousands of Ω m. Molten rocks have low resistivities (1–0.1 Ω m) and in the presence of partial melting, the effective resistivity falls sharply by several orders of magnitude at constant temperature. Electromagnetic sounding with low frequency the magnetic variations will allow probing the electric structure of the Martian mantle, and thus provide information on its thermal structure and mineralogy.

The electrical conductivity profile obtained by the inversion of magnetic data will complement the seismic model of the crust and upper mantle in order to better constrain the composition, which can then be used to compute elastic properties and density profiles inside Mars (Olsen, 1999; Tarits, 2001, 2002; Langlais and Purucker, 2007).

2.2.5. *Heat flow*

The average heat lost from a planetary surface reflects, at the most basic level, its bulk composition in terms of its content of radiogenic elements and thus provides a key constraint on the composition of the material from which the planet formed.

Current estimates of paleo heat flow on Mars are derived from estimates of elastic lithosphere thickness, interior evolution models, models of the decay of radiogenic elements, and simple plate cooling models (McGovern et al., 2002, 2004; Montesi and Zuber, 2004; Neumann et al., 2004; Belleguic et al., 2005; Hoogenboom and Smrekar, 2006; Breuer and Spohn, 2006; Grott, 2009; Grott and Breuer, 2008a, 2008b, 2010; Morschhauser et al., 2011). However, large uncertainties are associated with all of the above approaches and many estimates are based on lithospheric thickness estimates at volcanoes, which may not be representative of the bulk of the planet.

Following accretion, heat flow has been estimated to be high, of the order of 100 mW/m^2, but dropped quickly to ~15–50 mW/m^2 during the later evolution (McGovern et al., 2002, 2004). Simple plate cooling models predict present day heat flow in the range of 5–15 mW/m^2 and serve as a lower limit to the plausible heat flow values expected for a chondritic Mars. Stagnant lid models predict somewhat higher values of 20–25 mW/m^2 (Hauck and Phillips, 2002; Grott and Breuer, 2009).

Loading of the lithosphere by the polar caps is generally assumed to be relatively recent and these regions have the only well-determined surface deflections for which the present day elastic thicknesses (T_e) have been estimated (Johnson et al., 2000; Phillips et al., 2008). The obtained values of $T_e > 300$ km indicate that heat flow at the north pole is reduced with respect to chondritic composition models (Wänke and Dreibus, 1994), with $T_e > 300$ km corresponding to surface heat flows below 13 mW m^{-2} (Grott and Breuer, 2010, also compare Fig. 1). This implies that Mars either has a bulk composition which is subchondritic in heat producing elements, or that the surface heat flow is spatially heterogeneous and heat is transported through volcanically active regions like Tharsis instead (Grott and Breuer, 2009, 2010; Kiefer and Li, 2009). If an active plume underneath Tharsis does indeed exist, heat flow in this region would be expected to be elevated with values reaching 25–50 mW/m^2 (Grott and Breuer, 2010). Expected regional variations of surface heat flow are shown in Fig. 1.

2.2.6. Mars meteorology

Martian global atmospheric circulation is characterized by a dynamical regime similar to that of the Earth, since both planets are fast rotators (Leovy, 2001; Zurek and Martin, 1993). At solstices, the Hadley cell presents an ascending branch in the summer hemisphere and a descending branch in the winter hemisphere. Close to the equinoxes, two Hadley cells ascend from the equatorial region transporting heat towards both northern and southern mid-latitudes.

Martian circulation is modulated and modified by several kinds of atmospheric wave. Among these waves, the diurnal or semi-diurnal thermal tides, which are excited by the diurnal cycle of solar heating and propagate upwards with increasing amplitude, are of particular importance. The tidal waves are thought to be the primary phenomena controlling the dynamics above altitudes of 50 km. The traveling planetary waves observed at high and mid-latitudes in winter are of primary interest since they affect the dust and condensable species transport; they may also explain the zonal asymmetry observed in dust loading and water vapor abundance.

The general circulation is intimately related to the cycles of H_2O, CO_2 and dust. Indeed it influences the surface fluxes of dust and condensable species, their transport around the planet, and the heat advection over the Polar Regions. In turn, the presence of dust and clouds affects the energy balance of the atmosphere, and thus the thermal structure and atmospheric dynamics. Dust is the major atmospheric heating agent, whereas water ice particles also affect the planetary radiation budget through infrared cooling.

The observation of Martian weather by several surface stations will expand the existing very limited information on surface climate and circulation (based primarily on the two Viking Landers) to a much more comprehensive basis. Consistent time series of measurements of pressure, temperature, and wind will provide important new information on planetary scale circulation systems, on local and regional flows, and on the planetary boundary layer, and will provide key constraints on models of the response of the Martian atmosphere to variations in atmospheric mass, orbital parameters, and dust loading. Time series measurements can delineate diurnal and seasonal variations, as well as irregular variations related to storm systems on scales ranging from dust devils to global dust storms. Through the use of covariance analysis, time series from surface stations can be used to infer scale dependent properties of circulations at scales well below that of the station spacing. A limited number of stations deployed with care concerning their relative locations can be much more valuable, therefore, than the number of stations alone would suggest. Surface station measurements are also needed to define the vertical fluxes of momentum, heat, and water vapor in the surface boundary layer. These are critical forcing factors for the general circulation and water cycle.

Surface wind measurements have been performed by the Viking landers, Pathfinder, and Phoenix, although precise measurements of wind speed to date are still only available from the Viking Landers as Pathfinder and Phoenix were only equipped with relatively crude 'tell-tale' wind vanes (Gunnlaugsson et al., 2008; Holstein-Rathlou et al., 2010). Currently, wind patterns are generally derived from the temperature field using the thermal gradient wind approximation or more sophisticated techniques such as data assimilation (e.g., Lewis et al., 2007). However, such techniques suffer from inaccuracies on Mars because of the near-surface winds driven by the strong diurnal cycle, the large amplitude of the waves above 40 km and the difficulty of accounting for complex wave-mean interactions. Earth-based single-dish and interferometric millimetric observations have demonstrated the feasibility of tracking the Doppler shift of CO-lines to measure high-altitude winds. There are only a few such measurements, but they suggest that retrograde winds dominate around 60-km-altitude at almost all latitudes, even during equinox (Moreno et al., 2009). This strongly disagrees with thermal wind estimates based on MGS/TES data as well as with theoretical GCM predictions and urgently needs verification and with the report of prograde winds at high southern latitudes (Sonnabend et al., 2006).

3. Science objectives

3.1. Internal structure

3.1.1. Crustal thickness

The crustal thickness predictions of plate tectonics and stagnant lid models overlap somewhat given the wide range of parameters (initial mantle temperature, timing of initiation of plate tectonics, etc.). SEIS will provide absolute tie-points for crustal thickness that, when combined with heat flow measurements from HP^3, allows discrimination among models of crustal formation.

3.1.2. Mantle transition phases

For the Earth, it is known that olivine–spinel and spinel–perovskite transitions occur in the mantle. For Mars, it may be that a spinel–perovskite transition also occurs. Recent calculations suggest that these phase transitions may have important implications for the convection flow field in the mantle. SEIS will provide the depths of phase transitions that, when combined with heat flow measurements from HP^3 and core dimension measurements from RISE, allows discrimination among models of internal dynamics and the thermal evolution.

3.1.3. Core state and dimension

The factor of 10 improvement in core size determination provided by RISE from nutations (Dehant et al., 2009, 2011) reduces uncertainty in core size from 500 km (with no geochemical assumptions) to 50 km (or less, with additional constraints from SEIS), providing much tighter constraints on heat flow from the core (via constraints on the global temperature profile), core composition, and thermal evolution models for the core and mantle.

3.2. Mineralogy

At the coarsest scale the core is an iron rich alloy, the mantle consists of silicate rock, and the crust consists mainly of basaltic rock. Density increases within these layers mostly as a consequence of compression in response to the increasing pressure and of denser materials in the deep Earth, but there may also be mineralogical boundaries. SEIS will provide an averaged profile of seismic velocities that, when combined with induced magnetic field measurements from MAG and core dimension measurements from RISE, allows determination of the temperature and composition of Mars' interior, along with the density and electrical conductivity. These in turn provide constraints on the mineralogy inside Mars. These measurements will allow the determination of the depth of material discontinuities, the seismic velocities and attenuation, the density and the electrical conductivity. For the first time, interior models of Mars constraining the mineralogy, temperature, and physical state (including positions of the interfaces) of the planet can be derived from this future rich data set.

3.3. Interior thermal structure

The details of planetary thermal evolution and internal convection are governed by the amount of heat acquired during accretion, the abundance of radioactive isotopes, the degree of differentiation, and the nature of convective processes, as explained above. The surface heat flow is an important parameter that provides insight into the internal heat budget and dynamics. This can be measured with temperature sensors at different depths in the Martian soil. HP^3 can constrain the heat flow to within 10%.

Together with measurements of the thermo-physical properties of the soil, the long term monitoring of the temperature–depth profile from HP^3 will allow the determination of the surface planetary heat flow.

The outermost, rheologically stiff, layer of the planet (or lithosphere) has a base which is a defined by an isotherm or by a constant value of the homologous temperature (the ratio of temperature and melting temperature).

The temperature gradient in the lithosphere is significantly steeper than the gradients in most of the underlying mantle and the core as a consequence of the dominance of conductive heat transfer. In the sub-lithosphere mantle, heat is transferred by subsolidus convection and the temperature follows the adiabatic gradient through most of the sub-lithosphere mantle. Superadiabatic gradients in the sub-lithosphere mantle occur in thermal boundary layers at the bottom and top of the mantle and in regions of phase transformation. The mantle consists of a mixture of minerals and is separated into distinct pressure and temperature dependent mineralogies and phase change boundaries. SEIS will provide averaged profiles of seismic velocities that, when combined with heat flow measurements from HP^3, with magnetic field measurements, and with core dimension measurements from RISE, allow for a determination of the thermal structure.

3.4. Geological activity

SEIS will determine the seismicity of and impact frequency on Mars and HP^3 will produce the first measurement of the thermal state of the subsurface. These provide a fundamental measure of the geologic vitality of the planet and a direct measure of the current cratering rate, and point to regions for more intensive study by orbital or landed instruments. The level of seismicity gives a measure of the contemporary level of tectonic and perhaps volcanic activity, both in terms of intensity and geographic distribution. Current estimates of seismicity depend on thermal calculations (Phillips et al., 1990) or extrapolation of historical faulting (Golombek et al., 1992; Knapmeyer et al., 2006). Thus the measurement of seismicity, regardless of the actual number of events detected, provides fundamental information about the dynamics of Mars.

The distribution of quake epicenters may be highly non-uniform, as it reflects the current state of stress of the crust. The observed population of faults on Mars (Anderson et al., 2001, 2008; Knapmeyer et al., 2006) exhibits globally distributed compressional strain, with the notable exception of the Thesis rise where plume-driven volcanism and extension dominate. Based on crater counts, it has been suggested (e.g., Lucchitta, 1987; Neukum et al., 2004) that the Tharsis and Elysium volcanoes might still be active today. Ongoing volcanic and perhaps uplift activity will result in seismic activity concentrated in the Tharsis and Elysium rises. However, even the simplistic assumptions that the existing faults constitute zones of weakness where thermoelastic stress is released predominantly and that quakes are equally likely on all of these faults, result in a concentration of seismicity in the Tharsis region (Knapmeyer et al., 2006). Observation of a local seismicity level exceeding that predicted by this hypothesis, or observation of events with the specific characteristics of volcanic tremor, would thus confirm the existence of volcanic or dynamically driven tectonic activity on Mars today.

The unique characteristics of impact seismograms, characterized by a relatively low cutoff frequency compared to quakes (Lognonné et al., 2009; Gudkova et al., 2011) allow them to be distinguished from endogenic events (marsquakes). These seismograms can thus provide a direct measure of the current rate of impacts.

The thermal state of the planet is a key determinant for all endogenic geologic processes. Volcanism and tectonic deformation derive their driving energy directly from the heat engine of the interior. The thermal gradient also determines the thickness of the elastic lithosphere and the depth of partial melting, which controls magma generation. An estimate of the interior heat flow from using the HP^3 data eliminates a major uncertainty in predicting the depth of the liquid water stability zone on Mars.

3.5. Implications of improved constraints on early evolution

3.5.1. Dynamo and remanent crustal magnetization

A better determination of core size, density, temperature and mineralogy leads to improved understanding of core history. For example, current MOI estimates indicate that at least part of the core is liquid (Yoder et al., 2003; Balmino et al., 2006; Konopliv et al., 2006, 2011; Marty et al., 2009; Smith et al., 2009), but do not eliminate the hypothesis that rapid solidification of most of the core lead to dynamo shut down (Stevenson, 2001). This would only happen if the mantle is cold and the fraction of light elements very small, which is unlikely. Tighter bounds on the MOI and rotation and orientation observation will help refute this hypothesis. SEIS measurements, heat flow measurements from HP^3, magnetic field measurements from MAG, and improved core dimension measurements from RISE, could thus provide constraints on the early dynamo.

Tighter constraints on heat flow through time are important for understanding the vertical distribution of crustal magnetization. The very high depth-integrated magnetization values inferred from the magnetic anomalies imply much higher levels of magnetization than is typical of terrestrial rocks, very thick layers of magnetized crust, or both (Arkani-Hamed, 2004; Langlais et al., 2004). Some studies suggest magnetization extending deep into the crust (50–70 km) (Frawley and Taylor, 2004), while others find good agreement with shallower magnetized bodies (Dunlop and Arkani-Hamed, 2005). For typical Curie temperatures of $\sim 500\,^{\circ}C$, these results are puzzling, given that the dynamo is thought to have been active early, when heat flow must have been high.

3.5.2. Habitability and water

Early Mars had higher heat flow and at least transient periods of liquid water, including standing water on the surface. This was likely the most hospitable time period on Mars (Solomon et al., 2005), with heat flow providing an abundant potential energy source for biological processes. Little is known about the history of water at that time. Water in the shallow crust may have been a result of outgassing due to crustal formation (Hauck and Phillips, 2002; Morschhauser et al., 2011), magmatism (Jakosky and

Phillips, 2001), or simply higher heat flow (Postawko and Fanale, 1993). Additionally, water may have been stabilized by the presence of a magnetic field that prevented solar wind removal of hydrogen from the upper atmosphere. Thus early thermal evolution is key to understanding the role of water in Mars' geologically, and possibly biologically, most active time period. SEIS measurements, heat flow measurements from HP^3, and magnetic field measurements from MAG, thus provide constraints on the existence of liquid water at the surface in the past and in the sub-surface of Mars at present. The present-day heat flow also constrains the depth of the cryosphere and the region where liquid water might currently be present.

3.6. Implications for meteorology

Meteorology measurements (temperature, pressure, wind, humidity) provide constraints on the existence of liquid water at the surface and in the sub-surface of Mars both at present and in the past. The signature of most meteorological phenomena can be analyzed in the temperature structure and from direct measurements of wind and surface pressure. The general circulation can be constrained from temperature and pressure measurements of sufficient precision and frequency, especially if such measurements can be correlated with simultaneous measurements at more than one location.

Individual lander measurements of wind and temperature are especially useful for characterizing the planetary boundary layer, which is a critical region of the atmosphere that determines the exchange of heat, momentum, moisture, dust and chemical constituents between surface and atmosphere. Lander data are very important for this region of the atmosphere, particularly if such measurements have sufficient precision and time resolution to infer turbulent fluxes of momentum and heat. There have been relatively few measurements made on Mars of such quality so far, yet the Martian planetary boundary layer exhibits a more extreme range of conditions than found on Earth. So simple extrapolations of terrestrial parameterizations of boundary layer transport are not well verified at the present time, especially under conditions of strongly stable or unstable stratification.

Dust storms (regional or global) observed in concert with lower atmospheric measurements (temperatures and winds), will expand our understanding not only of the lower and middle atmosphere but also of the global thermospheric responses and associated radiative/dynamical processes operating on short time scales.

3.7. Seismology

SEIS is the critical instrument for delineating the deep interior structure of Mars, including the thickness and structure of the crust, the composition and structure of the mantle and the core. The power of seismology derives from the amount of information contained in a seismic signal. The ground vibrations detected by a seismometer reflect the characteristics of the original source, the geometry of the path taken from the source to the receiver (and thus the structure of the planet) and the physical properties of the material through which it has passed.

We can classify seismic analysis techniques into two types: the first does not require knowledge of the source location and the other is source location dependent. The first type of analysis can therefore be performed on measurements from a single station, while the other needs multiple landers, typically a network of three or more stations.

Traditional source location analysis (often associated with a network approach) is based on arrival times of body waves acquired by a widely distributed network of stations. Locating a quake requires therefore at least four seismic travel times (for determining its 3-D location and time), while the location of surface impacts associated with meteorites requires three seismic travel times. By using this method with the measurement of P and S wave arrival times only, the retrieval of structure information requires the third station for quakes and the second station for surface impacts. For both cases the detection of secondary phases or surface waves, while providing additional constraints on the structure, does not significantly change the capacity to locate sources. Azimuth data (i.e. the measurement of the initial ground vibration direction with a three-axis seismometer) significantly improve the three- and two-station cases, as they enable two stations to locate the epicenter (assuming spherically symmetric interior structure), leaving the arrival time information available for the determination of the depth, time and radial velocity. These azimuth data are however sensitive to lateral variations and to seismic noise and require at minimum a carefully installed three-axis instrument.

Over the past few decades however a wide variety of analysis techniques have been developed to extract information about the properties of the Earth's interior and seismic sources themselves using the data acquired from only one or two seismometers. In addition, such configurations can return interior seismic information:

(i) For seismic source with known location, e.g., impact craters located by orbital imaging and artificial impacts. If the impact time is known, both the P and S arrival time can be used on any seismometer, while differential travel times can be used when the time is not known.
(ii) For large quakes generating signals at long distances or for globally distributed seismic sources generating continuous seismic background. In these two cases, auto-correlation (with one station) and cross-correlation (with two stations) provide the free oscillation frequencies, the surface wave phase/group velocities, or the surface wave Greens function, any of which can be directly inverted for interior models.

In the following sections we outline several specific approaches to determine the seismicity, crustal structure and deep interior structure using data from one or two seismic stations. The science return of a network mission with many stations has been covered in previous publications (see references above).

3.7.1. Mars' seismicity

The distribution of seismic activity is determined by monitoring the teleseismic body wave frequency band ($\sim$0.1–2.5 Hz) for seismic events. The efficiency of detection will be mainly related to the quality of the instrument and installation and to the frequency bandwidth of the records. The approximate epicentral distances of seismic events are derived from the differential P–S arrival time on the vertical record. At the beginning of the mission, the location error is $\sim$10% (reflecting the range of a-priori estimates of Mars velocity models). The refinement of heat flow and crustal thickness (from HP^3 and SEIS, respectively) will produce better constraints on the upper mantle temperature. When added to SNC constraints on upper mantle mineralogy, significant improvements will be possible in the a priori bounds of event locations and, consequently, in upper mantle seismic velocities.

The lack of detection of seismicity by the Viking experiment is consistent with an upper estimate of Martian activity comparable to the Earth's intraplate activity (Anderson et al., 1977; Goins and Lazarewicz, 1979) with a total moment release midway between the Earth and Moon (Golombek et al., 1992). Theoretical estimates from thermoelastic cooling (Phillips et al., 2001) and estimates of seismic moment release from observed surface faults (Golombek et al., 1992) predict a level of activity within this bound but still ~100 times greater than the observed shallow moonquake activity (as would be expected for the more geologically active Mars). This level would provide ~50 quakes of seismic moment $\geq 10^{15}$ N m (a globally detectable quake, roughly equivalent to terrestrial magnitude $m_b=4$) per (Earth) year (Gutenberg, 1945a, 1945b). There should be ~5 times more quakes for each unit decrease in moment magnitude (or a factor of 30 decrease in seismic moment) (Knapmeyer et al., 2006).

Meteorite impacts provide an additional source of seismic events for analysis (e.g., Davis, 1993). On the Moon, impacts constitute ~20% of all observed events, and a similarly large number are expected for Mars: for the same impactor mass, although the atmospheric entry velocity is two times smaller (with possible further atmospheric deceleration for the smallest events), the frequency of impacts is 2–4 times larger. The Apollo 14 seismometer detected about 100 events per year generating ground velocity larger than 10^{-9} m/s and 10 per year with ground velocity larger than 10^{-8} m/s (Lognonné and Johnson, 2007), well within the detection capabilities of SEIS (see below). These provide a "guaranteed" set of seismic events independent of the level of tectonic activity. Fig. 4 shows a simulation of the expected distribution of impacts during one Martian year. The ablation of the meteorites by the atmosphere is integrated using the method of Lognonné and Johnson (2007) and the distribution of impacts is calculated by the method of Le Feuvre and Wieczorek (2008, 2011) and Lognonné et al. (2009). These results show that a few 10 s of impacts equivalent or larger to the Lunar Excursion Module (LEM) impact on the Moon will be detected per Martian year and might potentially be located if high-resolution cameras are contemporaneous operational on orbiting Mars spacecraft. Smaller impacts can be detected and located with meter-resolution instruments, and in total, modeling suggests that the number of impacts detectable on Mars by a seismometer with a sensitivity better than the expected seismic noise level (~10^{-9} m/s^2 at 0.5 Hz, see below), should be a few hundred per year (Davis, 1993; Lognonné and Johnson, 2007). Initial processing of the SEIS data could provide an approximate azimuth and range for such events. These approximate locations can be used to target the resulting new craters on the surface with high-resolution orbital imaging. If these fresh craters are found, the seismograms will then be constrained in terms of epicentral distance, and P–S differential travel times can be used for seismic inversions of the crust and upper mantle structure.

3.7.2. Signal to noise expectations

With the lack of seismic data, only an estimate of the signal to noise ratio can be made. With no ocean (the major source of terrestrial noise between 0.07 and 0.14 Hz, with amplitudes typically greater than ~10^{-8} m/s^2), we expect relatively low seismic noise. Estimates for the ground accelerations produced by local wind pressure fluctuation have amplitudes of the order of 10^{-9} m/s^2 in the range of 0.1–0.01 Hz with wind speed of the order of 4 m/s (Lognonné and Mosser, 1993). Moreover, terrestrial tests have shown that the direct effects of wind on a seismometer can be decreased to similar levels when protected by a light windshield (Lognonné et al., 1996) and decorrelated from atmospheric data (e.g., Beauduin et al., 1996; Montagner et al., 1998).

The transmission properties of Mars are obviously unknown. However, we can estimate these properties by analogy with the Earth, tempered by experience from Apollo measurements on the Moon. The greatest uncertainties are in attenuation (Q) and scattering. Seismic Q can be extrapolated from the observed tidal Q (Smith and Born, 1976; Lainey et al., 2007) or calculated for modeled mantle compositions and temperatures. Diffusive scattering dominates seismic records from the Moon (Toksöz et al., 1974), and significantly reduces the ability to detect seismic phases. However, strong scattering is not likely to be a factor on Mars as very low intrinsic attenuation, thought to be possible

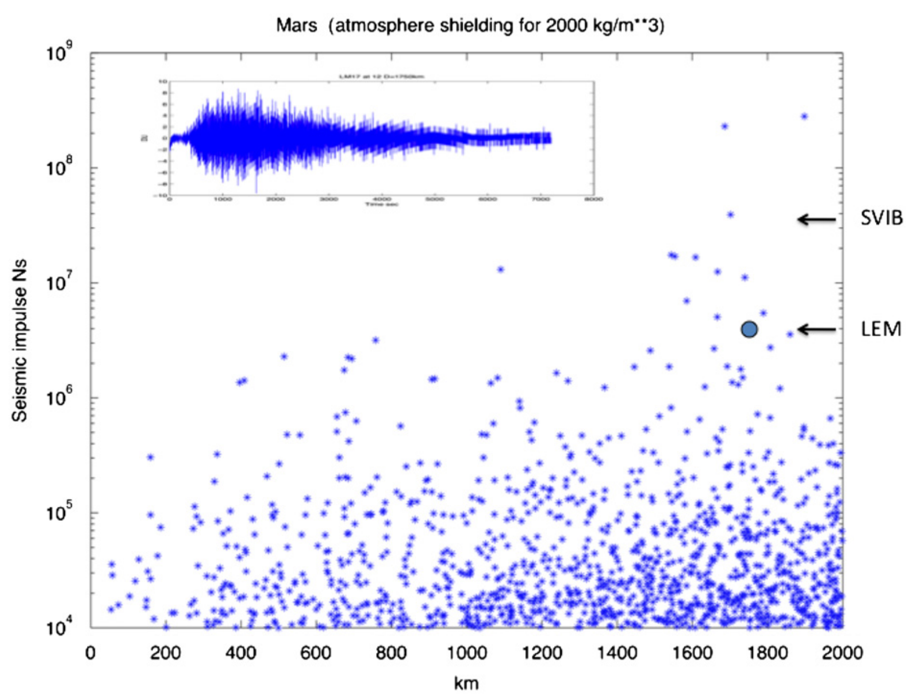

Fig. 4. Simulation of impacts during one Martian year at a given distance from a station. The amplitude of the impacts is the seismic impulse, roughly proportional to mv, where m is the final impact mass and v is the impact velocity. The value of the lunar seismic impulses from Saturn V upper stage and Lunar Module artificial impacts are given by arrows on the seismic impulse axis (see Lognonné et al., 2009, for details and values). The seismogram corresponding to the impact of Apollo 17 Lunar Module recorded at Apollo 12 station (epicentral distance of 1750 km) is shown at the top and indicated by the filled circle on the Distance-Seismic impulse plot. Impacts larger than Apollo Lunar Module impulses will generate craters of about 5 m in diameter, which could be easily detected from orbit.

only under extreme low-volatile conditions in hard vacuum, is required. It should be noted that the effects of both Q and scattering are less pronounced at the lower frequencies ($<\sim$0.1 Hz) enabled by modern VBB (very-broad-band) seismometer technology.

These estimates have been used to compute the expected amplitudes of body waves (Mocquet, 1999) and surface waves (Lognonné et al., 1996) and are summarized in Fig. 5. The calculations suggest that P and S waves from a quake of 10^{15} N m can be detected globally with SNR > 5 for a sensitivity of 10^{-9} m/s^2, while surface waves at 0.01 Hz will be similarly detectable for moment larger than 10^{17} N m.

From the inferred level of activity and well-understood partitioning of energy as a function of frequency in seismic signals, the required sensitivity of a Mars seismometer can be derived. This sensitivity can be formally represented as an equivalent noise floor, with dimensions of m/s^2/Hz$^{1/2}$. In order to take full advantage of the techniques given the expected seismic environment, the required performance of a seismometer should be no worse than 10^{-9} m/s^2/Hz$^{1/2}$ in the frequency range 0.1–1 Hz, and if possible comparable to the performance of the Apollo seismometer ($\sim 10^{-11}$ m/s^2/Hz$^{1/2}$) in order to extract all available seismic information.

3.7.3. Single station body wave processing

3.7.3.1. Source azimuth and distance. The polarization of the initial P arrival gives the direction to the source of the P wave. This is measured using the horizontal components, yielding an error of $<10^\circ$ in azimuth for conservative projected levels of horizontal component noise, resulting in a $\sim$15% uncertainty in transverse location similar to the radial uncertainty from P–S analysis. Thus events can be roughly localized within a $\sim$100 km uncertainty at distances of 1000 km. With the distance and location known approximately, the spatial distribution and magnitude of seismicity (from amplitude and the approximate distance) can be determined. This is a fundamental parameter of the seismic environment of a planet.

3.7.3.2. Mantle and core reflectors. As noted above, the joint determination of P and S arrival times provides estimates of epicentral distance to about 15%, and RISE reduces the uncertainty in core radius to 20–50 km. With these constraints on the ray paths we can use the refined interior structure models with synthetic seismogram analysis to identify later-arriving phases (Fig. 6) The additional differential measurements of arrival times, such as PcP–P, PcS–S and ScS–S, as well as comparison of their relative amplitudes to P, provide additional constraints on the seismic velocities and attenuation in the deep mantle of Mars. These constraints help refine the core size estimate and place bounds on lower mantle discontinuities. Amplitude measurements of these phases as a function of frequency can also provide the first measurements of attenuation in the deep mantle, a key parameter for the optimization of any future seismic network mission.

3.7.3.3. Crustal thickness determination. Estimates of the crustal thickness are possible using the receiver function method. When a P or S wave strikes a discontinuity in a planet, it generates reflected and transmitted waves of both P and S. Because of this, waves from distant earthquakes passing through a layered medium such as the crust or upper mantle can generate complicated seismograms containing many echoes. Such seismograms can be processed to generate simplified waveforms called receiver functions (Phinney, 1964; Langston, 1979). These can be inverted to yield the variation of shear velocity with depth, and they are particularly sensitive to strong velocity discontinuities. By back-projecting the receiver functions into the interior, the depths and even lateral variations of distinct boundaries such as the Moho can be reconstructed (e.g.,

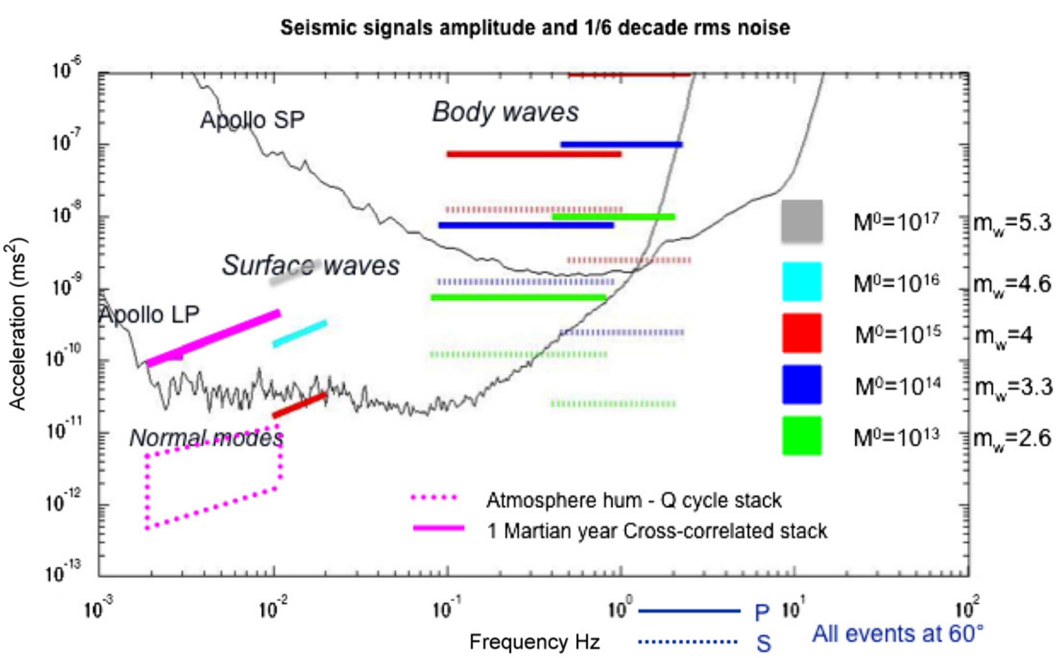

Fig. 5. Expected seismic signal strength on Mars for various sizes of events, compared to the noise floor of the Apollo seismometers. Noise levels are those of the Apollo instruments (black lines, LP is the long-period instrument and SP the short-period instrument) and are, for each frequency *f*, defined as the rms amplitude for a 1/6 decade bandwidth (roughly $f \pm 20\% f$). Body wave signal amplitudes are extrapolated from Mocquet (1999) for two bandwidths, 0.1–1 Hz and 0.5–2.5 Hz, and are represented by a different color for several seismic moments, all for an epicentral distance of 60°. Surface wave amplitudes are determined for a 3 mHz seismic hum due to atmospheric forcing (see Lognonné et al., 1998; Lognonné and Johnson, 2007). All amplitudes are those recorded by a single station, except the 3 mHz seismic hum amplitude, for which the expected amplitude of the cross-correlated signal from two stations is given. Units of seismic moment are N m. (For interpretation of the figure legend, the reader is referred to the web version of this article.)

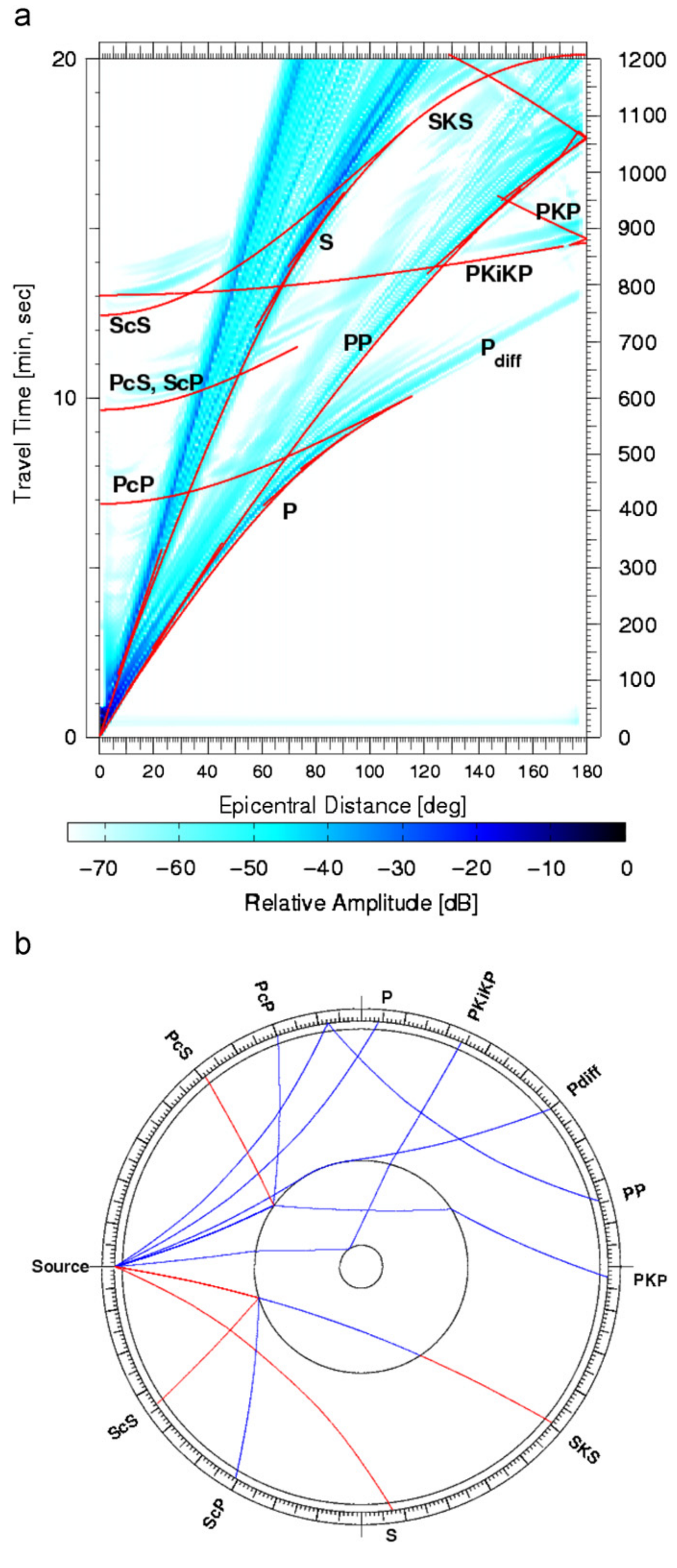

Fig. 6. (a) This synthetic seismogram section depicting recorded ground velocity (color coded) as function of epicentral distance (horizontal) and time (vertical) , shows the wealth of later arrivals ("phases") from different ray paths that can be used by SEIS to constrain mantle and core structure. Blue shades denote amplitudes relative to those at the epicenter, red lines are ray theoretical arrival times. Computed for model A of Sohl and Spohn (1997) with an additional 300 km inner core and a source depth of 10 km. (b) Example ray paths of the phases shown in (a). Blue segments denote P wave propagation, red segments S wave propagation. Computed for the same Mars model as (a). (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

Knapmeyer and Harjes, 2000). Crustal thickness and Poisson's ratio of the crust can be determined independent of the velocity structure determination (Zhu and Kanamori, 2000), yielding an anchor for gravity-based crustal thickness mapping. Receiver functions are a powerful tool for studying the depths to the crust–mantle boundary or to other layering within the crust, and are computed from single seismograms without requiring source location or time. This method has been widely used on the Earth and was successfully applied to the Moon (Vinnik et al., 2001).

3.7.4. Surface wave processing

3.7.4.1. Surface wave dispersion. Surface waves are low-frequency seismic waves that propagate in the crust and upper mantle and owe their existence to the presence of the free surface. By sampling the crust and upper-mantle, surface waves are an important source of information. The depths to which surface waves are sensitive depend on frequency, with low-frequency waves "feeling" to greater depths and therefore propagating at higher speeds. This results in dispersion, with low-frequency waves arriving earlier than higher frequencies. The details of the relation between frequency and group velocity are directly relatable to subsurface structure and provide moreover information on lateral crustal variations (e.g., Larmat et al., 2008). They are extremely sensitive to the crustal thickness, as shown by Fig. 7, and variations $\geq 10\%$ are typical for crustal variations of 20 km. The sensitivity to the upper mantle is also important, as the group velocity of surface waves (or the differential group velocity between the fundamental and the overtones) varies by 5–10% for models with different iron content (Mocquet et al., 1996).

In order to obtain velocity from arrival time, an estimate of the distance from the source is necessary. This can be obtained from the P–S arrival time difference or more precisely from the $R1$–$R2$ difference ($R1$ is the direct Rayleigh wave arrival, whereas $R2$ is the arrival of the wave propagating around the planet in the opposite direction).

3.7.5. Normal modes

For a pair of seismometers, the most effective techniques for studying deep structure use normal mode frequencies, which do not require knowledge of the source location. Normal mode spectral peaks from 5 to 20 mHz (the frequency range sensitive to mantle structure) should be identifiable for a detection noise level of 10^{-9} m/s^2/Hz$^{1/2}$ (Lognonné et al., 1996; Lognonné and Johnson, 2007). This can be accomplished by seismogram analysis of a large quake of moment $\geq 10^{18}$ N m (equivalent $m_b \sim 6$; Fig. 8). The likelihood of such a quake occurring during a restricted mission lifetime such as one Martian year is estimated to be

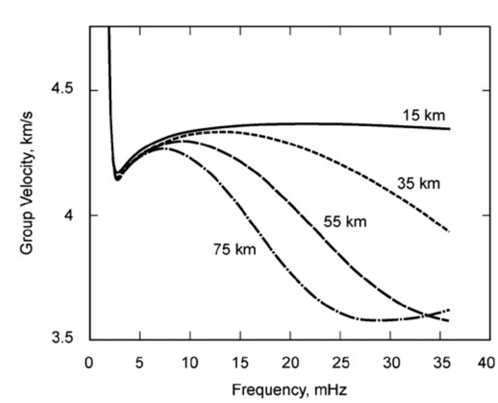

Fig. 7. The dispersion of Rayleigh waves is extremely sensitive to crustal thickness. Model B of Sohl and Spohn (1997) is used for crustal velocity and density.

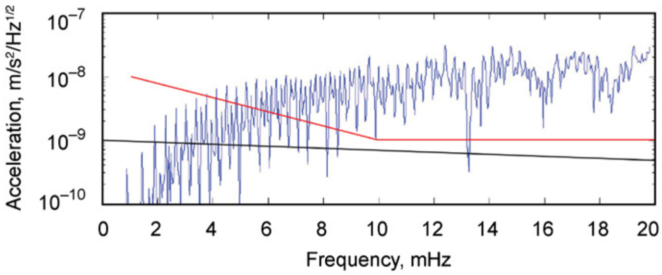

Fig. 8. The detectability of normal modes for large quakes is shown in this plot of the synthetic acceleration amplitude spectrum for a quake with a seismic moment of 10^{18} N m (Lognonné and Johnson, 2007). The black and red lines show the SEIS expected and required sensitivity, respectively.

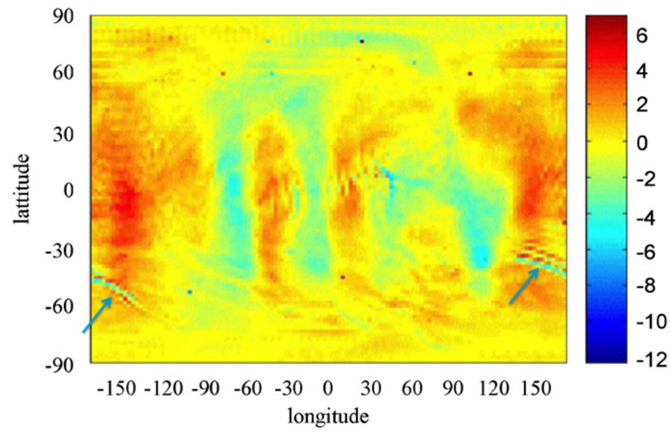

Fig. 9. Snapshot of the acceleration field from a Martian GCM. Note the effects of an atmospheric front in the southern hemisphere (indicated by arrows), providing a highly localized excitation, and of the global-scale pressure field associated with daily variations (reprinted from Lognonné and Johnson, 2007).

low (Golombek et al., 1992, Knapmeyer et al., 2006), but one can get the same results at frequencies higher than about 10 mHz, by stacking multiple quakes with an equivalent cumulative moment (for instance ten equivalent $m_b=5$ earthquakes).

Similar techniques can be applied to the background noise generated globally by atmospheric dynamics (Kobayashi and Nishida, 1998a, 1998b), the so-called seismic "hum". Calculations based on excitation by turbulence in the boundary layer that do not take into account resonance effects, non-turbulent wind and pressure variations associated with atmospheric circulation (Tanimoto, 1999, 2001) yield amplitudes for Mars $\sim$0.1 nanogal, a factor of 2–3 smaller than on Earth.

A more precise estimate of excitation takes into account atmospheric pressure variations and winds (Lognonné and Johnson, 2007). The relative contributions of the oceans and atmosphere are still debated for the Earth, but on Mars this hum is generated by a dynamic coupling of the normal modes with the atmospheric circulation. This coupling occurs with measurable amplitudes at angular order ≤ 10, at much lower frequencies than on the Earth ($\sim$29). Thus the coupling coefficient is comparable to that on Earth at angular orders of about 10 and typically a factor 10 smaller for angular orders greater than 20 (as a consequence of the smaller atmospheric density) (Lognonné and Johnson, 2007). But the generally lower coupling on Mars is offset somewhat by the significantly larger temperature fluctuations and winds.

Global circulation models (Forget et al., 1999) can be used with these coupling coefficients to produce an estimate of the continuous excitation of normal modes (Fig. 9). Significant excitations are observed, including atmospheric fronts in the southern hemisphere. Such models can be integrated over time to provide mode amplitudes, and result in amplitudes 5–10 times smaller than observed on Earth. These estimates are likely a lower bound, as turbulence is not included. In addition, the fact that Martian circulation is more coherent than the Earth's on a daily scale provides the possibility of increasing the SNR by long-duration stacking. When combined with a GCM, joint measurements of pressure variations and seismic hum can provide unique insight into the dynamic coupling between the atmosphere and the surface of a planet devoid of oceans. These observations in turn enable a better understanding of hum generation on the Earth.

3.7.6. Phobos tides

The size of the Martian core can be constrained through measurements by SEIS of the solid tide induced by Phobos, which gives a combination of the Love numbers h_2 and k_2 (Lognonné and Mosser, 1993; Van Hoolst et al., 2003). The Phobos tides, which produce a ground acceleration signal of order 0.5×10^{-8} m/s^2, are sub-diurnal with typical periods shorter than 6 h (Van Hoolst et al., 2003). They are thus below the primary seismic frequency range and provide a unique link between high frequency (seismic) and ultra-low frequency (geodetic) observations of Mars' interior. The SEIS measurement is essentially limited by the temperature noise ($\sim$0.5 K rms in a bandwidth of 1 mHz around the Phobos orbital frequency [Van Hoolst et al., 2003]). Insulation and temperature decorrelation post-processing can reduce the effective noise by a factor of 100–250, leading to a final noise of about 10^{-10} m/s^2 rms after one Mars year of stacking. This noise amplitude results in a core radius determination error of <60 km.

3.8. Geodesy

RISE uses the intrinsic precision of the Deep Space Antennas (NASA Deep Space Network - DSN or ESA TRACKing station—ESTRACK) and the spacecraft communications system to derive key information about the deepest structure of the mantle and core. The RISE investigation infers interior structure from its effect on variations in the orientation of Mars' rotation axis with respect to inertial space. The precession, nutation, and polar motion of Mars result from the interaction of the interior mass distribution with the gravity of the Sun and the angular momentum of the atmosphere. RISE provides improved estimates of these motions by analyzing the radio link from the lander with the spacecraft and/or the Earth. The projected improvements (e.g., a factor of 10 in precession) over existing measurements from Viking, MPF, and the MGS-Mars Odyssey-Mars Reconnaissance Orbiter (MRO) orbiters result from the increased total time span between Viking and a future lander mission, the longer contiguous time span (1 Mars year vs. 90 sols for MPF) and better tracking accuracy (particularly with respect to Viking).

Precision tracking of the Martian surface is performed through radio links between stations on the Earth and the lander on Mars. The experiment uses the X-band or Ka-band communications transponder to obtain periodic two-way Doppler and ranging measurements from the radio link. Additional interferometry measurements using the two radio links together can further increase the precision on the measurements. These measurements accumulated over a long period of time can be used to obtain Mars' rotation behavior (precession, nutations, and length of day (LOD) variations). Precession (long-term changes in the rotational orientation) and nutations (periodic changes in the rotational orientation) as well as polar motion (motion of the planet's surface with respect to its rotation axis) are measured and used to obtain information about Mars' interior. At the same time, measurement of variations in Mars' rotation rate reveals variations of the MOI due to seasonal mass transfer between the atmosphere and polar caps (Cazenave and Balmino, 1981; Chao and Rubincam, 1990; Yoder and Standish, 1997; Folkner et al., 1997; Defraigne et al., 2000; Dehant et al., 2006, 2009, 2011; Van den Acker et al., 2002; Sanchez et al., 2004; Karatekin et al., 2005, 2006a, 2006b; Zuber et al., 2007).

The geophysical requirements on tracking are derived from technical analysis (Asmar et al., 2005; Iess et al., 2003, 2009; Tortora et al., 2002, 2004) and detailed numerical simulations of science parameter extraction. In order to provide the tracking precision necessary for the RISE investigation, the lander must transpond the carrier signal from Earth with an end-to-end stability, usually represented by the Allan Standard Deviation, of at least 10^{-13} over a 1000 s integration time, equivalent to $\sim$0.1 mm/s, over time scales of 60 s with equivalent isotropic radiated power greater than 6 W. The lander must be tracked for at least 60 min per week for no less than 12 months. The DSN or the ESTRACK antennas can be used to transmit and receive the X-band (or Ka-band) carrier signal and measure the Doppler shift. Further, use of signals from two landers enables interferometric measurements and results in a further increase in the precision of the radioscience measurements and thus of the geophysical parameters.

3.8.1. *Precession and nutations*

Precession measurements (from RISE), together with measurements of the nutations (from RISE) and tides (from SEIS), improve the determination of the MOI of the whole planet and thus the radius of the core. For a specific interior composition or range of possible compositions, the core radius is expected to be determined with a precision of a few tens of km (compared to ± 200 km currently). A precise measurement of variations in the spin axis also enables an independent (and more precise) determination of the size of the core via core resonances in nutation amplitudes. The amplification of this resonance depends on the size, moments of inertia (MOI) of the core and the planet, and flattening of the core. For a large core, the amplification can be very large, ensuring the detection of the free core nutation and determination of the core MOI (Dehant et al., 2000a, 2000b; Defraigne et al., 2003).

A large inner core can also have a measurable effect on nutations. Due to the existence of another resonance (the free inner core nutation), there would be amplification in the prograde band of the nutation frequencies, which would result in the cancellation of the largest prograde semi-annual liquid core nutation (Defraigne et al., 2003; Van Hoolst et al., 2000a, 2000b; Dehant et al., 2003). Failure to detect the amplification of the semi-annual nutation together with the confirmation of a liquid outer core from the retrograde nutation band or from k_2 would be strong evidence for a large solid inner core.

3.8.2. *Polar motion*

Polar motion is the motion of the planet with respect to its rotation axis or equivalently the motion of the mean rotation axis with respect to the figure axis in a frame tied to Mars. It is induced by atmospheric motion and is related to the angular momentum exchange between the atmosphere and the solid planet as for the LOD variations. A full quarter of the atmosphere is involved in the CO_2 sublimation/condensation cycle between the polar caps. The components of polar motion are thus dominantly at seasonal timescales. Additionally, as the rotation axis is not necessarily coincident with the figure axis, there might be a wobble of Mars. This is the so-called Chandler Wobble, which is at a period of about 200 days (Van Hoolst et al., 2000a, 2000b). As the atmosphere behavior is not purely harmonic and noise may arise, the Chandler Wobble contribution to polar motion might be at the level of a few meters, depending on the dissipation within the planet (Q).

The observation of polar motion from radio links between a lander at the surface of Mars and the Earth is only possible for landers at high latitudes; the contributions to the Doppler shift from an equatorial lander are too small to be observable (see Le Maistre et al., 2010), except for interferometry methods.

3.8.3. *Length-of-day*

The rate of Mars' rotation around its axis is not uniform due to angular momentum exchange between the solid planet and the atmosphere. There is a variation in the LOD which is, at the seasonal timescale, due to the sublimation and condensation process of CO_2 already mentioned. A lander at the surface of Mars on the equator undergoes a huge excursion of its position with respect to a constant rotation of up to 15 m peak to peak over a Martian year.

This phenomenon can be computed from a GCM. The angular momentum is transferred to the solid planet by three kinds of coupling mechanisms: (1) the pressure torque related to the atmospheric pressure on the topography, (2) the gravitational torque related to the mass anomalies inside Mars and in the atmosphere, and (3) the friction torque related to the wind friction on the Martian surface (see Karatekin et al., 2011). The angular momentum of the atmosphere consists of two parts: the mass term related to the rigid rotation of the atmosphere with the solid Mars, directly involving the surface pressure all over the surface; and the motion term related to the relative angular momentum of the atmosphere, directly involving winds.

Present day GCMs allow computation of the seasonal changes induced in the rotation of Mars as well as in the low-degree gravity coefficients. But there are still unexplained differences between the computations and the observations. Additionally, inter-seasonal variability is expected in the time variation of the low-degree gravity coefficients due to dust storm contributions, which is not perfectly in agreement between observations and model simulations from GCMs.

3.9. *Heat flow measurements*

HP^3 will perform thermal measurements in the shallow subsurface that uniquely constrain the surface planetary heat flow. The technique used to measure Martian heat flow (q) uses Fourier's law, $q=\delta(\Delta T/\Delta z)$, where $\Delta T/\Delta z$ is the vertical thermal gradient and δ is the thermal conductivity. Thus the heat flow determination requires measurements of both the thermal gradient and the thermal conductivity at the same location. These measurements should be taken at a depth where the diurnal and preferably also the annual thermal cycles are attenuated. However, shallower measurements are admissible if they span the period of at least one complete annual cycle (Grott et al., 2007).

Long term climate-related surface temperature variations (e.g., Christensen et al., 2003; Mustard et al., 2001; Kreslavsky and Head, 2002; Head et al., 2003; Helbert et al., 2005) could potentially disturb near-subsurface temperatures. However, the effects from changes of the Martian spin/orbit parameters have been modeled by Grott et al. (2007) and were found to have relatively minor effects on surface heat flow at low latitudes, with maximum perturbations being smaller than 15% for periods applicable to the Martian obliquity cycle (Laskar et al., 2004). Furthermore, this signal can be modeled with some confidence and is minimized by choosing a near-equatorial landing site (Mellon and Jakosky, 1992).

Although heat flow on Mars is unlikely to produce the large difference measured between continents and ocean basins on Earth, some geographical variations depending on such factors as the enrichment of radioactive elements and crustal thickness are expected (Grott and Breuer, 2010). Measurements of the surface heat flow in three to four geologically representative regions (e.g., the northern lowlands, southern highlands, Tharsis, and Hellas) will provide firm constraints on the average heat loss from the planet.

Other considerations which need to be taken into account include the possibility of active mantle plumes underneath the Tharsis and Elysium volcanic provinces. Crater counts suggest that some areas of Tharsis may have been volcanically active within the last tens of millions of years and if plumes indeed exist underneath these regions, heat flow would be expected to be elevated there. A plume signature would be clearly visible in even a single surface heat flow measurement, as heat flow is expected to be elevated by a factor of two with respect to the background chondritic model in the presence of active plumes (Grott and Breuer, 2010).

It will not be necessary to conduct heat flow measurements simultaneously, and these measurements therefore do not require a network to be operating. The measurements made by one or two landers could thus constitute a baseline and provide an important first sample of the surface heat flow, which can be built upon by subsequent lander missions.

For a present day heat flux of approximately 20 mW/m^2 and a thermal conductivity of 0.025 W/mK, the expected thermal gradient is $\sim$0.8 K/m. Thus for a reliable determination of the thermal gradient ($\sim$5%), we require that the accuracy of the individual temperature measurements over a 2-m depth interval (see below) should be better than 0.05 K. The observed temperature profile need not be vertical as long as the relative depth of the temperature sensors is known to within 2 cm (Spohn et al., 2010).

Orbital thermal inertia measurements provide an estimate of the thermal properties of the upper few cm of the regolith (Christensen et al., 2003), and for the low rock abundance areas of interest for a future lander mission the thermal inertia is expected to be of order 250 $J/m^2\, s^{1/2}\, K$. This translates to a diffusivity at the surface of 0.05 mm^2/s, or a skin depth of 4 cm for the diurnal and 3 m for the annual wave. HP^3 is designed to reach depths of 3–5 m in order to get below the penetration depth of the annual wave, but even if the full deployment depth is not reached, annual perturbations can be modeled and removed if measurements are taken over the period of at least a full Martian year (Grott et al., 2007).

Other thermal perturbations which need to be taken into account include shadows cast by the lander and airbags (Grott, 2009), and these perturbations can be minimized by deploying the heat flow probe 1–3 m away from the shadow source. Additionally, shadowing can be modeled if measurements are extended over the period of a full Martian year (Grott, 2009), relaxing the requirements on instrument deployment.

During mole action, regolith at the tip at the mole is radially displaced, resulting in soil compaction and an associated increase of the thermal conductivity as the added pressure increases the contact area between grains. For Apollo, regolith compaction was identified as the most likely cause for raised thermal conductivity values (Grott et al., 2010), but the amount of soil compaction is expected to be considerably less using the mole system on Mars, because no borestem maintaining overpressure after mole passage will be used. Furthermore, the reduced role of grain contact area on thermal conductivity due to the atmosphere and the lower relative density of the Martian soil will act to reduce the influence of soil compaction caused by the passage of the mole.

3.10. Magnetism measurements (orbiter and landers)

Magnetic field measurements have never been performed at the Martian surface. Such measurements would thus provide the first characterization of the magnetic field and magnetic activity at the surface of Mars. In addition, it is a primary objective to provide information on the internal structure by electromagnetic sounding, and, if possible, to measure the secular variation of the magnetic field. Surface magnetometers deployed from landers could provide a substantial contribution to crustal induced field investigations by performing measurements over a long period of time.

During MGS aerobraking, measurements were made down to an altitude of about 100 km, revealing large amplitude, small scale crustal magnetic anomalies, dominantly over the southern hemisphere, indicative of remanent magnetization. These low altitude data were completed by higher altitude measurements, at about 400-km, until November 2006 when MGS ceased operation. This low altitude magnetic survey is unfortunately very sparse, although these are likely the most valuable measurements. On Earth there are crustal field structures of scale much smaller than O(100) km wavelengths resolvable by MGS. For example the horizontal scale length associated with seafloor spreading and a reversing dynamo is of the order of 10 km, and that of most other structural features and mineral deposits is smaller. Resolving structures of such a spatial scale is well beyond the reach of orbiting spacecraft, but magnetic measurements made during the decent phase of the landers would really complement MGS existing measurements, especially if the landing target was near a magnetic anomaly. It would then be possible to:

- Determine quantitative limits on the depth, volume and magnetization intensity for the crustal magnetic anomalies measured during the MGS aerobraking phase.
- Determine a possible correlation between the obtained high-resolution magnetic anomaly features and surface geology and geophysical properties.
- Test existing models for crustal formation and alteration.
- Support interpretation of surface mineralogy data obtained by other in-situ and remote sensing instruments.

Even a single point measurement performed by a lander after descent would to some extent contribute to these objectives.

Synergy with possible magnetic measurements from an orbiter will benefit both the crustal field mapping and the electromagnetic sounding of the crust and mantle:

- The ground-based data will help separate spatial and time variations observed from orbit.
- Through the ground-based data, specifically "magnetically quiet" periods can be singled out for investigation for crustal fields, as is customary for the Earth.
- Complementary orbiter measurements will help characterize the geometry of the signal from inducing currents at the surface, data which is necessary for electromagnetic sounding.

Measurements performed from orbit or from magnetometers even far from magnetic anomalies will provide the temporal and seasonal characteristics of the magnetic induced field and will probe the interior of Mars by allowing the determination of impedance and the conductivity profiles of the planet in synergy with the lander measurements.

3.11. Meteorology

Meteorology measurements provide present constraints on GCMs for better understanding the CO_2, water and dust cycles. The observation of Martian weather at surface locations will greatly enhance the limited information currently available on Mars' climate and circulation which, for surface measurements, is still based primarily on the two Viking landers. Consistent time series measurements of pressure, temperature, and wind will provide important new information on planetary scale circulation systems, on local and regional flows, and on the planetary boundary layer. This information will place constraints on models of the response of the Martian atmosphere to variations in atmospheric mass, orbital parameters, and dust loading. Time series measurements can delineate diurnal and seasonal variations, as well as irregular variations related to storm systems on scales ranging from dust devils to global dust storms. Time series measurements from surface stations can be used to infer scale-dependent properties of circulations at scales well below that of the station spacing. Surface station measurements are also needed to define the vertical fluxes of momentum, heat, and water vapor in the surface boundary layer. These are critical forcing factors for the general circulation and water cycle.

Surface pressure should ideally bemeasured with an accuracy of 3 Pa on a typical mean surface pressure of 610 Pa. Variations on a roughly daily timescale can reach up to 30 Pa just due to traveling weather systems. Shorter timescale variations from dust devils may exceed that. Seasonal changes are of order 200 Pa. A future lander mission will record all of these phenomena throughout the diurnal cycle and for a full Mars year to reveal the traveling weather systems, the thermal tides, the seasonal trends and possibly even catch some dust devils. Time-averaged air temperatures just above the lander should be recorded with a precision of 0.1 K and an absolute accuracy of ± 1 K. These measurements can be used to understand the effects of Martian weather systems, including large-scale dust storms. Near-surface wind speed should bemeasured just above the lander deck with an accuracy of at least 0.1 m/s or better. Even if direction is not measured, wind speed alone provides a critical indicator of dust lifting, and hence the connections between the pressure and temperature signals, and the associated winds can help increase our understanding of dust storms on Mars. Correlating the winds with pressure and temperature yields information about the nature of the weather systems themselves (e.g., Barnes 1980, 1981).

As discussed in Section 2.2.6 winds have not been observed on Mars by any spacecraft other than the Viking landers, Pathfinder and Phoenix, and wind speed was only measured with reasonable accuracy from the Viking Landers. Until now wind patterns have been mostly derived from the temperature field using the thermal gradient wind approximation or via more sophisticated techniques such as data assimilation. But these techniques are only as good as the thermal data used as input, which are mainly derived from remote sensing measurements that suffer from limited coverage and vertical resolution, especially close to the ground. The uncertainty introduced by these remote measurements suggests that major findings in atmospheric dynamics may be best obtained by simultaneous 3-D measurements of temperature and wind from orbit, although some surface information is really needed to tie down the recovery of wind and surface pressure. The value of surface station measurements will therefore be greatly enhanced if made in conjunction with orbiter measurements of the 3-D temperature and wind distributions, completed by dust and water vapor distributions.

Lander measurements can be used to characterize the distribution of dust in the atmosphere, search for and characterize atmospheric clouds, and search for and characterize dust devils. Images of the sky near the sun provide a direct measurement of atmospheric dust scattering, and the acquisition of images at different times of day can provide information about the vertical dust distribution. Direct measurements of the sky brightness also provide a better understanding of the diffuse illumination conditions at the landing site, important for interpreting surface texture. Cloud images provide detailed information about their morphology, and time-lapse images yield information about cloud formation and evolution during the day.

Far-field surveys can provide direct imaging of dust devils at the local landing site. Time-lapse images of these dust devils provide information about their formation, evolution, speed, and duration. This information is used to calculate dust devil volume and density, key parameters for modeling global atmospheric dust loading and parameterizing dust lifting processes in GCMs.

Lander missions with seismometers onboard will be equipped with atmospheric pressure, air temperature and wind speed sensors for seismic noise decorrelation, which is crucial for the interpretation of seismic data. This relatively simple but valuable meteorological package can provide critical data to further our understanding of the meteorological processes acting near the surface of Mars as well as for planning future surface operations at Mars. While we have basic near-surface results from four locations on Mars (at Viking, MPF, MER and Phoenix landing sites), Martian meteorological phenomena are still not fully understood. With the breadth of regional environmental conditions on Mars, a new lander mission provides a key addition to this body of meteorological data.

Meteorological measurements during EDL provide important input for Mars meteorology as well. Accelerometers on such a lander provide crucial atmospheric structure data for designing future vehicles that will transit Mars' atmosphere. Improved measurements of the density structure at altitudes above 50 km are needed for reliable predictions of aerobraking and aerocapture maneuvers. Profiles of density, pressure, and temperature between the thermosphere and the surface were made by Viking, MPF, MER and Phoenix (Schofield et al., 1997; Seiff and Kirk, 1977; Magalhães et al., 1999; Withers, 2009), and in situ measurements of the upper atmosphere were derived from aerobraking deceleration measurements of MGS (Bougher et al., 1999) and MRO. But profiles at different seasons, solar time and geographic location are needed.

3.12. Geology

The surface geology investigation characterizes the geology of the future landing sites, and provides "ground truth" for orbital remote sensing data.

Stereo panoramic imaging provides the basic data set used to establish the broad geologic evolution of the local region and tie these observations with orbital remote sensing data during the landing site selection process (e.g., Christensen and Moore, 1992; Golombek et al., 2008). Imaging at the Viking, Mars Pathfinder (MPF), Mars Exploration Rovers (MER), and Phoenix sites established the general geologic setting, identified the geologic materials present, and quantified their areal coverage. These images allowed the classification of different soil types at the sites, the areal coverage and size-frequency distribution of rocks and outcrop, the types and amount of eolian materials, and the morphology of craters, all of which can be used to better understand the evolution of the landscape.

Previous in-situ measurements of the rock abundance, types and coverage of soils (and their physical properties), thermal inertia, albedo, and topographic slope have all agreed with orbital remote sensing estimates and show that the materials at the landing sites can be used as "ground truth" for the materials that make up most of the equatorial and mid-latitude regions of Mars. By understanding the materials that make up the landing sites, the sites can be also be used as "ground truth" for orbital remote sensing data of Mars and thus aid in future landing site selection (Golombek et al., 2008).

4. Conclusions

The pioneering measurements provided by a geophysical lander mission will be interpreted using a broad array of sophisticated analysis techniques that take full advantage of the richness of these fundamental geophysical data sets, and of their synergetic use. Geophysical questions can still be addressed while using a single station. Nevertheless, more landers provide an enormous increase in the scientific yield. In particular, there is a significant step in considering at least two stations.

Concerning orbiter results, Mars Global Surveyor, Mars Odyssey, Mars Express, and Mars Reconnaissance Orbiter have completely revamped our understanding of the Mars' geological evolution, in conjunction with the ground truth provided by the Mars Exploration Rovers and the Phoenix mission. Major advances have been made, such as discovering water-ice below the surface, mapping of the various types of ice in the polar regions, establishing the history of water abundance on the surface of Mars in view of the minerals formed at different epochs, the presence of methane in the atmosphere, mid-latitude auroras above crustal magnetic fields, and much younger timescales for volcanism and glacial processes. Indeed, the potential presence of methane suggests that either volcanism or biological processes could currently be active on Mars.

Much remains to be done in several key areas. Some major objectives for investigations at Mars relate to the structure of Mars' interior and the escape of the atmosphere and its response to solar activity (including the escape of neutral species, for which no data currently exist). Our knowledge of Mars' atmospheric dynamics also remains low, particularly in regards to coupled low-middle-high atmosphere circulation, the water cycle, chemical cycles and role of electricity. The link between magnetic and mineralogical signatures and the history of the atmosphere provides an additional and ongoing opportunity for investigation. The NASA MAVEN spacecraft (due for launch in 2013) and the ExoMars Trace Gas Orbiter (due for launch in 2016) address some of these questions, emphasizing aeronomy and trace gas measurements, respectively. The emphasis for the NASA Mars Science Laboratory (MSL, due for launch in 2011) and for the combined ESA-NASA ExoMars/MAX-C Rover is the search for indicators of life and habitability. A single geophysical station in the time frame 2016–2018, either as a stand-alone mission (Banerdt et al., 2010; Elkins-Tanton et al., 2011) and/or added on to the Mars 2018 Sky-crane mission (Braun et al., 2011) followed by a multiple lander mission in the time frame 2020–2022, will be unique in addressing fundamental global questions, which can only be answered by landers. This will moreover provide opportunities for synergy between Europe and the US (obviously a first step in this direction has been done for the ExoMars and Trace Gas Orbiter missions), possibly with other countries (Japan, China, or Russia, etc.).

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