A view of extraterrestrial soils

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Summary

The nature of soils on celestial bodies other than Earth is a growing area of research in planetary geology. However, disagreement over the significance of these deposits arises in part due to the lack of a unified concept and definition of soil in the literature. The pragmatic definition “medium for plant growth” is taken by some to imply the necessity of biota for soil to exist, and has been commonly adopted in the planetary science community. In contrast, a more complex and informative definition bases on scientific theory: soil is the (bio)geochemically/physically altered material at the surface of a planetary body that encompasses surficial extraterrestrial telluric deposits. This definition emphasizes that soil is a body that retains information about its environmental history and that does not need the presence of life to form. Four decades of missions have gathered geochemical information of the surface of planets and bodies within the Solar System, and more information is quickly increasing. Reviewing the current knowledge on properties of extraterrestrial regoliths, we conclude that the surficial deposits of Venus, Mars, and our moon should be considered soils in a pedological sense, and that Mercury and some large asteroids are mantled by possible soil candidates. A key environmental distinction between Earth and other Solar System bodies is the presence of life, and due to this dissimilarity in soil forming processes, it is plausible to distinguish these (presently) abiotic soils as Astrosols. Attempts to provide detailed classifications of extraterrestrial soils are premature given our poor current knowledge of the Universe, but they highlight the fact that Earth possesses nearly abiotic environments that lend themselves to understanding more about telluric bodies of the Solar System.

He found himself in the neighbourhood of the asteroids 325, 326, 327, 328, 329, and 330. He began, therefore, by visiting them, in order to add to his knowledge.

(Excerpted from the “The Little Prince” by Antoine de Saint-Exupéry)

Introduction

At 22,17 on July 20, 1969 Neil Armstrong set foot on the Moon (Figure 1) and declared: «One small step for man, one giant leap for mankind». Since then, the “lunar regolith” has been studied in detail on samples collected during that first landing, and from samples retrieved on subsequent manned or unmanned missions. A comparable database is accumulating for the Martian regolith, and lesser amounts of information exist for the surface of other bodies in Solar System.

The four inner planets, including Earth, are telluric (rocky) planets depleted in volatile elements. There is a broad similarity in their lithologies, along with enormous differences in atmospheric chemistry and surface temperatures. While the regolith on the Moon and Mars has been referred to as soil, the definition applied to these materials indirectly implies a distinction from soil in the earth sense: “The term soil is used here to denote any loose, unconsolidated materials that can be distinguished from rocks, bedrock, or strongly cohesive sediments. No implication for the presence or absence of organic materials or living matter is intended” (Soderblom *et al.*, 2004). Here, we argue that this definition is inconsistent with a scientific definition of soil that most earth scientists use.

The surfaces of celestial bodies are strikingly inhospitable to the forms of life we know on Earth (*e.g.* Hornek, 2008; Weber & Greenberg, 1985). Therefore, does the presence of life represent a distinguishing *condicio sine qua non* between terrestrial and extraterrestrial soils?

In this paper, we discuss the definition of soil, and review the soil forming environments on rocky bodies of the Solar System in order to consider the distribution of soil types on such bodies.

Definition of soil

The definition of soil hinges on human perception and concepts, and it has been stated “soil can be what you want it to be” (Arnold, 2006, p. 2). While that view is applicable to non-scientists, the scientific definition of soil is generally agreed upon and is in wide use. Soil is a natural, historical body at a planet’s surface that is a partial record of the environmental conditions that have existed at that location. The Soil Taxonomy (Soil Survey Staff, 2006, p. 1) defines soil as “a natural body comprised of solids (minerals and organic matter), liquid, and gases that occurs on the land surface, occupies space, and is characterised by one or both of the following: horizons, or layers, that are distinguishable from the initial material as a result of additions, losses, transfers, and transformations of energy and matter *or* the ability to support rooted plants in a natural environment”. This definition clearly illustrates that while soils *may* grow plants, they do *not* need plants, or processes driven by plants, to be soils. This is particularly important on Earth, in regions such as Antarctica and Chile, where ancient soils have formed entirely without the presence of higher plants due to cold and/or aridity. While it has been demonstrated that some extraterrestrial materials are able to sustain plant growth on the Earth (Mautner, 1997), soil is the *in situ* veneer of a planetary surface – not a potting mixture or other horticultural materials. Soils are historical entities that retain information on their climatic and geochemical history.

Soils form in response to the well known soil forming factors. How will these factors differ in extraterrestrial settings? Based on our Earth perspective, rocky or solid parent materials are required as a substrate. Climates, e.g. the environmental conditions of a planetary surface, will be far different, and drive different chemical and physical processes, than on Earth. The near absence of O_2, the lack of an ozone layer and/or atmospheres, will all greatly change the expected nature of soil processes and alteration products. Additionally, in tectonically inactive and stable surfaces, the amount of time available for soil processes may be orders of magnitude longer than on Earth. Finally, based on our present knowledge, all extraterrestrial soils will be abiotic, but the necessity of biota for soils is debatable (*e.g.* Markevitz, 1997). While some authors have strictly associated soil with biological activity (*e.g.* Richter & Markevitz, 1995), this restrictive view, of course, excludes important parts of our own planet from pedological study – *e.g.* the Dry Valleys of Antarctica (Ugolini & Anderson, 1973) or the Atacama Desert of Chile (Ewing *et al.*, 2006; Navarro-González *et al.*, 2008) – regions (Figure 2) that have virtually life-free soils with advanced horizonation. Additionally, very young deposits can lack both plants and horizon differentiation; however, they are soils that are simply very young (*e.g.* Ugolini and Edmonds, 1983; Amundson, 2006). Therefore, the USDA Soil Taxonomy definition, quoted above, includes these types of soils on Earth, and serves as a template for the consideration of soils on other planetary bodies. Soil is the skin of Earth, a skin weathered by contact with the atmosphere. Planetary bodies have a chemical nature and atmospheric conditions that can be similar or very dissimilar to those on Earth. Below, we examine these in more detail.

Soil processes on other planets

Because of the enormous difference in atmospheres on the inner planets and moons, the processes that modify the planetary surfaces can vary greatly from what we are familiar with on Earth. Foremost, chemical and physical processes on most extraterrestrial bodies in the Solar System are limited by the lack of water. However, there are several polar solvents that could replace water, such as sulphuric, hydrofluoric, and hydrocyanic acids, and ammonia, methanol, and hydrazine (Schulze-Makuch & Irwin, 2006). Secondly, the most powerful proton acceptor on Earth, oxygen, is not present as free O_2 on nearby planetary bodies, but can be available in ozone (O_3) and CO_2. On rocky bodies, oxygen is abundant in tightly bound oxides such as SiO_2, Fe_2O_3, MgO, and Al_2O_3, but these require very high-energy processes to reduce. Even without water and molecular O, a variety of energy sources drive chemical reactions: thermal, osmotic, and ionic gradients, solar wind, magnetospheric energy, and radioactivity are among the most important (Schulze-Makuch & Irwin, 2006). Volcanic activity and meteorites supply both energy and material to planetary bodies.

Solar wind is a stream of high-energy electrons and protons that are able to escape star's gravity and, eventually settle on the surface of planetary bodies. Mineral grains interact with the atoms accelerated and ionised by shock waves. The energies of these ions, which depend on the shock velocity, are of the order of keV. Chemical changes were observed in the composition of olivine irradiated with relatively few keV H^+ and He^+ ions (Dukes *et al.* 1999). Demyk *et al.* (2001) caused the amorphisation of crystalline olivine, and its changes toward a pyroxene-type composition through selective loss in O and Mg, by irradiating dust with He^+ ions at energies of 4 and 10 keV and fluxes varying from $5\ 10^{16}$ to 10^{18} ions cm^{-2}. The thickness of the amorphised region was 40±15 nm and 90±10 nm for the 4 keV and 10 keV experiments, respectively. The amorphisation of the olivine also implied a porosity increase due to the formation of bubbles.

Meteorite impacts can liberate enough energy to destroy not only the meteorite, but part of the planetary surface, converting rock and minerals into gas at temperatures of several thousand kelvin. Some of the gas condenses to form tiny metal spheres surrounding the impact crater. The ejecta that falls back around the crater forms a rim that rises several metres above the surrounding plain and, depending on the size of the impact, up to around ten kilometres away. The rocks in the centre of the crater are intensely crushed and metamorphosised, often leading to formation of new “shocked” minerals.

Volcanic activity is an internal weathering force that supplies fresh minerals as either airborne solid ejecta or lava flows. However, all magmas release dissolved gases during eruptive episodes and/or between them, and these gases can then form acids that weather local or regional surficial deposits. Energy is released as explosions and tremors, which break rocks and alters minerals through heating.

The rest of this paper will examine what is known of extraterrestrial regoliths and the nature of their soils.

Nature of planetary parent materials and bodies

Soil, as we conceive it based on our experience on Earth, requires a solid parent material, which in the Solar System occurs only on the four inner planets: Mercury, Venus, Earth, and Mars. All of these have roughly the same structure: a central metallic core, mostly iron, with a surrounding silicate mantle. Telluric planets have canyons, craters, mountains, and volcanoes (Baker, 2008), and possess secondary atmospheres (generated through internal volcanism or comet impacts). Moons and asteroids are solid and have a silicate or a carbonaceous composition, but they lack both an iron core and an atmosphere. Telluric planets, some moons, asteroids, and transient comets of the Solar System are the only space bodies that we have direct observations from manned or unmanned landings.

Mercury

Mercury, the closest planet to the Sun, was first orbited by the NASA Mariner 10, in 1974, which obtained about 10,000 images, about 57% of the planet’s surface. In January 2008, the

NASA Messenger began sending bright, detailed images of Mercury’s cratered surface, included the hemisphere that Mariner 10 did not succeed in imaging.

The surface of Mercury is characterised by a variety of terrains ranging from heavily cratered to nearly crater free (Figure 3). Although volcanoes have not been yet identified, the smooth plains on Mercury indicate that molten material in the past filled low-lying areas. Mercury almost completely lacks an atmosphere that protects it from meteorites and solar wind. The surface temperature ranges between 100 and 740 K. The regoliths of Mercury are very dark, chiefly formed from anorthosite probably high in iron (McCord & Clark, 1979), finely dispersed products of impact crushing and thermoclastism. Microscopic processes of nanophase iron particle formation possibly can take place through repetitive dust impacts or through Ostwald ripening. The latter is a thermodynamically-driven process that occurs when a secondary phase precipitates out of a solid, and energetic factors cause the larger precipitates to grow, drawing material from the smaller precipitates, which consequently shrink (Sasaki & Kurahashi, 2004). The formation of glassy agglutinates due to meteorite impacts is also probable. Hapke (2001) identified solar wind sputtering and presence of submicroscopic metallic iron by deposition of vapours created by micrometeorite impact and vaporization as the main causes of surface weathering on Mercury. Loeffler *et al.* (2008) observed that olivine deposits produced reddening and darkening due to the production of nanoparticles of metallic iron. Coarse-grained regoliths are generally much less affected by many processes than fine-grained ones because space weathering is mostly a particle surface phenomenon that extends less than 1 μm into the exposed faces.

In the first of its three encounters with Mercury before settling into orbit around the planet in 2011, Messenger has captured a plethora of images of the surface that reveal that it has been greatly weathered by the impacts and solar bombardment (Solomon *et al.*, 2008). There is evidence of regolith maturation of Mercury’s surface, the most important being the occurrence of iron as nanoscale metal or oxide grains (McClintock *et al.*, 2008; Robinson *et al.*, 2008). Areas of intense reflectivity near vents seem to be pyroclastic deposits. Preliminary spectroscopic studies suggest a strange lack of ferrous iron from Mercury's crust and the probable presence of opaque minerals, one of which could be the secondary iron titanium oxide ilmenite (Melosh, 2008). Because the evidence available suggests that the Mercury surface is both physically and chemically altered, it would seem plausible to suggest that it possesses a soil mantle, one that has formed under conditions highly dissimilar to that on Earth.

Venus

Venus is sometimes called "Earth's sister planet" because of its similarity in both size and bulk composition. However, Venus' surface temperature is about 750 K, with small diurnal temperature fluctuations. This stable climate is caused by Venus' atmosphere, the densest of the four telluric planets, which imposes an atmospheric pressure 91 times that of sea level on Earth. Such a dense atmosphere suggests that the surface is protected from erosion by small cosmic particles. Venus' atmosphere consists mostly of carbon dioxide and sulphuric acid droplets, with traces of water and oxygen. Despite the occurrence of clouds, there is no precipitation. The planetary surface is characterised by mountains and highland areas surrounded by large, flat stony deserts (Figure 4). In 1975, the Russian probes Venera sent back the first close-up photographs of the planet's surface, which in places appeared covered with sharp-edged rocks or fine-grained dust. The flat stones, several tens of cm in size, seemed formed from outcrops as a consequence of thermal processes or corrosion carried out by the atmosphere. Other mechanisms considered capable of causing physical weathering include volcanic eruptions (some volcanoes may still be active today), impacts of large meteorites, and wind erosion (Surkov *et al.*, 1984). In 1982, Venera 13 landed on Venus, returning the first colour images and carrying out chemical analyses of the regolith. The sample analysed was found to contain Ca, Na, Si, Fe, Mg and K. It is hypothesised that the extremely acid atmosphere is effective at degrading rocks into secondary weathering products. The weathering of rocks on Venus should produce the end-products magnetite and anhydrite, given the environmental boundary conditions (Barsukov *et al.*, 1986), eventually making the soil on Venus unique in the Solar System.

Earth's Moon

The nearest and most studied extraterrestrial body is our moon. Moon is mantled by a layer of loose, heterogeneous fragmental debris that overlies solid rock. The Lunar regolith is unevenly distributed, varying in thickness from 3-5 meters on plains called "maria" (Latin for seas), to 10-20 meters in the highlands. Maria, which cover 85% of the total surface, are basalts, while the highland bedrocks are mostly composed of anorthosite. Since the end of the volcanic era, about 3 billion years ago, the lunar regolith has experienced the action of several weathering processes which include: physical reworking by meteorite impacts, deposition and chemical interactions of impact-generated and sputtered ions, irradiation by solar wind. As a result of these processes, lunar regolith shows horizonation, and several categories of centimetric and millimetric strata have been identified (Duke & Nagle, 1975). Some millimetric laminae are probably debris shed from adjacent rock fragments. Other dark, fine-grained, well-sorted thin laminae appear to be superficial zones reworked by micrometeorites. Rock fragments and large spatter agglutinates are commonly found in their original depositional orientation, or were reworked from the underlying units.

The NASA Apollo Programme enabled 12 astronauts to sample the Moon between 1969 and 1972 (Figure 5). From six landing sites almost 400 kg of rock fragments and fine earth were returned to Earth. The main observable chemical effect on the Lunar regolith is the formation of a 50-200 nm thick rim of weathering on grains. This rim differs from the core of grains, primarily in the relative proportions and oxidation states of the elements (Keller & McKay, 1997). For example olivine and pyroxenes have suffered significant loss of Mg, transformation of Fe^{2+} to nanophase Fe metal, and preferential sputtering of Ca and Al relative to Si and O. Rims on ilmenite grains are not amorphous as they are elsewhere, but are microcrystalline as a consequence of the preferential removal of Fe and O, along with a reduction of Fe^{2+} to Fe^{0} and Ti^{4+} to Ti^{3+}.

The solar wind hits the surface of the Moon unimpeded by magnetic fields or atmospheres. This wind darkens and reddens loose powders on the lunar surface in proportion to their Fe content (Hapke, 2001). Starukhina *et al.* (1999) estimated that condensation products of the clouds generated by sputtering and impacts should be distributed within a layer of regolith whose thickness is several times the diameter of the impactor. During thermal evaporation the rock-forming oxides are partially dissociated and the molecular oxygen is selectively lost due to its high volatility (Hapke *et al.*, 1975). Fractionation processes lead to formation of O-depleted, Fe-enriched films where the missing oxygen is compensated for by the reduction of the oxides with the lowest binding energy. It has been estimated that 30% of the metallic Fe is native to the lunar rocks, 30% is meteoritic in origin, and the remaining 40% is the result of reduction of FeO by space weathering (Morris, 1980).

A renewed interest in the Moon has led the NASA to schedule unmanned missions between 2008 and 2011 to map out landing sites and to find a spot for a permanent base (NASA, 2009a). Other agencies are planning missions explicitly focused to study the nature of the Lunar soils.

Mars

The remaining telluric planet, and the one that is farthest from the Sun, is Mars. Mars does not show any evidence of plate tectonics and active volcanism, hence its surface is relatively stable. However, a great difference in temperature between its two hemispheres induces strong winds, up to 400 km h^{-1}, which are able to carry powders from one side of the planet to the opposite one. The atmosphere is highly rarefied (average surface pressure is 1 Pa), consisting mostly (95%) of carbon dioxide and molecular nitrogen.

Even with the naked eye, the Fe oxide rich regolith of Mars is visible from Earth. Our knowledge of it has grown immeasurably due to 6 successful landing missions, all equipped with varying payloads of imaging, spectroscopic, and chemical instrumentation (Figure 6). The Martian regolith is a complex and spatially heterogeneous mixture of locally-derived volcanic rocks of basaltic composition partly pulverised by impacts, a variety of fluvial deposits, thick sequences of silicate and sulphate rich sedimentary rocks of possible marine origin, and a variety of aeolian features (*e.g.* Herkenhoff *et al.*, 2004; Grant *et al.*, 2004; Greeley *et al.*, 2004). The spacecrafts Viking 1 and 2 estimated that the Martian dust has bulk densities of 1.0-1.6 Mg m^{-3} and can comprise 70% of less than 2 mm particles at the immediate surface, most of which is less than 100 µm (Cherkasov & Shvarev, 1978).

The presence of olivine, which is an easily weatherable primary mineral, has been interpreted to mean that physical rather than chemical weathering processes currently dominate on Mars (Morris *et al.*, 2004). Extreme thermal oscillations are among the physical weathering processes. For example, the daytime surface temperature in equatorial regions of Mars reaches 300 K, but at night it drops to 180 K.

Mars has no sea, but has water in the form of ice close to the surface and adjacent to the permanently frozen poles, as revealed by NASA's Odyssey spacecraft and the Phoenix lander. The Martian landscape at high latitudes bears the imprint of particle sorting and surface patterns (Figure 6b) that are morphologically similar to patterned ground described on periglacial regions of Earth (*e.g.* Ugolini & Bockheim, 2008). A 4-centimeter deep “profile” (Figure 7), examined on June 1st, 2008 by the Phoenix lander, on the northern hemisphere of Mars at 68°, revealed evidence of subsurface ice. Water contents of Martian regolith range from <2% by weight to more than 60% (Mitrofanov *et al.*, 2002; Horneck, 2008). There is abundance of geomorphic and chemical evidence that liquid water was once active on Mars (*e.g.* Squyres *et al.*, 2008). Delta deposits, river terraces, thick sequences of sedimentary deposits rich in sulphates, and the growing identification of secondary phyllosilicates and hydrated secondary minerals all point to previous, and possibly periodic, aqueous chemical alteration of the planetary surface. Grey hematite, for example, is an iron-oxide indicative of past water, although it is not always associated with water. The reddish hue of Martian regolith is due to rusting iron minerals presumably formed a few billion years ago when Mars was warm and wet, but now that Mars is cold and dry, modern rusting may be due to a superoxide that forms on minerals exposed to ultraviolet rays in sunlight (Yen *et al.*, 2000). The NASA rover Opportunity, which landed on 2006 on Meridiani Planum, found hematite in concretions that formed in salt-laden deposits laid down via wind and/or water (Kerr, 2004). At “Hematite Slope”, Opportunity opened trenches about 50 centimetres long and 10 centimetres deep, discovering shiny round pebbles (“blueberries”) in fine-grained matrix. In the Martian regolith it was observed that “what's underneath is different than what's at the immediate surface” (Yen *et al.*, 2000), which is suggestive of soil horizonation.

Sulphates such as jarosite (potassium iron hydroxy sulphate) and kieserite (magnesium sulphate) appear to occur in many geological settings on the planet. Barrón *et al.* (2006) assessed that jarosite can transform directly to hematite in simulated Martian brines. Thick sedimentary sequences of salts have been identified, which on Earth form in standing water, possibly during evaporation (Vaniman *et al.*, 2004). However, sulphate and other salts are found intermingled with siliceous surficial deposits at multiple locations (Viking 1, Viking 2, Pathfinder, Spirit), associations that on Earth occur in hyperarid environments.

Burns (1987) proposed that on the young, volcanically active Mars, sulphuric acid derived from volcanic emissions would have mixed with any water and chemically eroded rock to produce a variety of sulphates, and would have contributed to pedogenesis. Other evidence that indicates the rocks have experienced aqueous alteration is the migration of chlorine and bromine into cracks and fissures on the surface of the rocks.

Clay minerals had been recently identified on Mars by orbiter-based spectroscopic analyses (Chevrier *et al.*, 2007). In particular, the OMEGA instrument onboard the ESA's Mars Express detected the presence of smectites (montmorillonite and nontronite) and chamosite (a member of the chlorite group) in various locations by visible-near-infrared hyperspectral reflectance imagery (Bibring *et al.*, 2005; Poulet *et al.*, 2005). Their presence indicates chemical weathering combined with leaching. Fe-rich phyllosilicates probably precipitated under weakly acidic to alkaline pH. It has been speculated that this process occurred early in Martian history, and was later followed by strongly acid weathering that led to sulphate deposits (Bibring *et al.*, 2006).

Amundson *et al.* (2008), based on comparisons to ancient soils in hyperarid regions of Earth, argued that soil at the Spirit and Pathfinder landing sites bear evidence of early stage chemical weathering and loss of major rock forming elements, and show a late stage addition of sulphate, chloride, and associated cations. This interpretation implies the accumulation of weathering products at depths below the Martian surface during time periods commonly considered being too dry for aqueous weathering.

Weathering products are found also on Martian meteorites that have been deposited on Earth. Taylor *et al.* (2002) attributed a round patch of smectite in the Martian meteorite Dhofar 019 to aqueous alteration on Mars. Similarly, Treiman *et al.* (1993) argued that the iddingsite (a fine-grained intergrowth of smectite clay, ferrihydrite and ionic salt) present in the Martian meteorite Lafayette might be a product of pre-terrestrial alteration. Equally important, Martian meteorites contain sulphates and carbonates with unique O isotope compositions that imply they were formed by oxidation in the Martian atmosphere (Farquhar & Thiemens, 2000) and were subsequently transported by aqueous carriers into the Martian regolith, prior to being ejected by a meteor impact.

In summary, the strong likelihood of an early wetter and warmer Mars, the evidence of depth related chemical features, and the evidence of recent or even on-going cryoturbation, all are indicative of a variety of soil types on Mars that likely have close Earth analogues.

Phobos and Deimos, the two small moons of Mars, were first visited in the 1970's by Mariner 9 and the two Vikings. Their importance in terms of human colonisation of space could be much greater than their size because they have been considered as potential staging areas for human colonies on Mars (Mautner, 1997; Veverka & Burns, 1980). Phobos and Deimos contain carbon-rich rock, but their low bulk densities imply they consist of a mixture of rock and ice. Both are heavily cratered. Phobos (Figure 8) is covered with a layer of fine dust about a meter thick, similar to the regolith on the Earth's Moon. Less is known about Deimos, but it is probable that it does not differ much from Phobos. Despite the parent rock composition of Phobos and Deimos is one of the best candidate to allow formation of Earth-like soils, at the present stage of knowledge it is impossible to speculate about the occurrence of soils on these moons.

Between Mars and the edge of the Solar System

Direct observations for the celestial bodies currently orbiting outside of Mars' orbit have not been made.

More than 100,000 asteroids lie in a belt between Mars and Jupiter. Asteroids are rocky bodies that are believed to be left over from the beginning of the Solar System. They have round or irregular shapes up to several hundred km across, but often are much smaller. Their small size deprives them of any internal heat and, hence, tectonics that could rejuvenate their surface. More than 75% of asteroids are C-type, namely they have a composition that is largely carbon-based and lacks hydrogen, helium and other gases. Analysis of density trends suggests that asteroids are divided into three general groups: asteroids that are essentially solid objects, asteroids with macroporosities around 20% that are probably heavily fractured, and asteroids with macroporosities over 30% that are loosely consolidated “rubble pile” structures (Britt *et al.*, 2002). While small asteroids have bare rock surfaces, the larger ones possess a sufficient gravitational force to allow the accumulation of regoliths, including a fraction of finer particles (Matson *et al.*, 1977). For example, the asteroid Vesta, ~500-600 km in diameter, has a gravity such that more than 95% of the ejecta from a meteoric impact falls back onto its surface. By scaling for differences in gravity, the regolith of the larger asteroids should resemble that of the Moon.

NEAR Shoemaker, the first spacecraft to soft-land on an asteroid, rendezvoused asteroid 433 Eros in 2001. Eros is approximately 13x13x33 km in size, the second largest near-Earth asteroid. NEAR Shoemaker was not designed as a lander and was deactivated after it arrived; hence, the last of a plethora of images of 433 Eros was taken from a range of 120 meters as the craft collided with the body. Eros' surface is dominated by a blanket of regolith littered by craters and boulders ranging in size from nearly 100 meters down to 8 meters. The regolith seems to have been darkened and reddened by the solar wind and micrometeorite impacts, while apparently fresh, bright material is confined to patches of fines set on a background of older fragmental debris (Trombka *et al.*, 2000). Most of the large boulders look relatively altered, or perhaps are stained by an adhering film of regolith particles.

The Japanese spacecraft Hayabusa was the first to take samples from an asteroid, the 535-m-long 25143 Itokawa, in 2005. The surface of 25143 Itokawa photographed by Hayabusa (Figure 9) appears bumpy, with knots of hard matter looking like lumps and only small amounts of powdery regolith. Data shows that the asteroid's density is very low (~1.9 Mg m^{-3}), suggesting Itokawa may be composed of small rocks held together by gravity (Abe *et al.*, 2006; Fujiwara *et al.*, 2006). Hayabusa did not land on the asteroid, but used a cone-shaped device that captured materials ejected by the impact of a metal pellet fired onto the surface (Okada *et al.*, 2006). Sample return is scheduled for 2010.

Io, the only telluric moon of Jupiter, was orbited by NASA Galileo spacecraft during the period 1996-2003. Io is heated by tidal forcing – exerted by the gravitational fields of Jupiter and two other moons, Europa and Ganymede – that causes volcanism. There is the possibility that Io started out similar in character to the other icy moons of Jupiter and lost its icy mantle over time due to tidal forcing. Most of Io's surface is 120 K or colder, despite the presence of volcanoes. The volcanism continuously reworks the ferric-sulphuric-silicatic crust (Simonelli *et al.*, 1997), thus reducing the time available for any pedogenic processes. However, the residence time of this crust, and the possible assortment of pedogenic processes operable, is unknown.

Titan (Figure 10a), the largest moon of Saturn, is so large that its gravity slightly compresses its interior. Titan is about half water ice and half rocky material, but is probably differentiated into several layers with a 3400 km rocky centre surrounded by several layers composed of different crystal forms of ice. Titan’s atmosphere is composed primarily of molecular nitrogen with 6% argon, a few percent methane, and trace amounts of at least a dozen of other organic compounds (i.e. ethane, hydrogen cyanide, carbon dioxide) and water. Titan possesses rivers and, apart from our planet, it is the only body known to possess lakes. However, the fluid is not water but methane and other liquefied hydrocarbons (Mitri *et al.*, 2007). A geologically varied surface, apparently modified by a mix of processes including strong fluvial erosion, impacts, and cryovolcanism was revealed by the NASA Cassini orbiter and the ESA Huygens probe that landed on Titan in 2005 (Elachi *et al.*, 2005). Cassini RADAR images show widespread plains covered by sand dunes probably composed of organic solids and water ice (Lorenz *et al.*, 2006).

Iapetus (Figure 10b) is the Saturn's third-largest moon. The most enigmatic feature of Iapetus is the roughly elliptical region centred on the leading side ("Cassini Region"), darker by a factor of about 10 than parts of the surrounding areas (Porco *et al.*, 2005). Such a feature and the Iapetus's asymmetry may perhaps be attributed to a thick primordial low-albedo subsurface layer of organics exhumed by impact erosion (Wilson & Sagan, 1996). At the current stage of knowledge, it is hard to hypothesise about the occurrence or nature of soils here and on Titan.

Comets and exoplanets

Comets are assemblages of ice and dust that periodically pass through the centre of the Solar System along highly elliptical orbits. When comets get close enough to the Sun, heat induces their partial sublimation so that jets of gas and dust form long tails that can be seen from Earth. Comets contain the most primitive material in the Solar System (*e.g.* Ishii *et al.*, 2008).

In 2005, the NASA Deep Impact space probe shot a projectile into comet Tempel 1 to obtain material to analyse. The first results showed that cloud contained more minerals and less ice than had been expected (OSIRIS Team, 2005). However, even if many comets do contain rocky debris, the dynamic nature of comets likely precludes the formation of any type of soil.

There is a rapidly growing number of stars in the universe that now have known planets orbiting them (Cole, 2001), and it is speculated that the number of planets in the universe exceeds the number of stars. These planets, called “exoplanets”, are being discovered continually and as of May 2009, 347 candidates had been identified (URL exoplanet.eu).

Recently, using the ESO 3.6-m telescope in La Silla, Chile, an Earth-like exoplanet, Gliese 581c, was discovered at 20.5 light years from the Earth, in the constellation of Libra (Udry *et al.*, 2007). Under the assumption that it is a rocky planet, Gliese 581c has a radius at least 50% larger than that of Earth, is about 5 times the mass of the Earth, and has a gravity approximately 2.2 times as strong as on Earth. It appears to be in the habitable zone of space surrounding the parent star red dwarf Gliese 581, where the surface temperatures might maintain liquid water. Gliese 581c may be tidally locked to its parent star, with one hemisphere always lit and the other always dark. Gliese 581c has a projected equilibrium surface temperature 270–310 K. However, the actual temperature on the surface depends on the planet's atmosphere, which remains unknown. In 2008, the NASA/ESA Hubble Space Telescope captured the first optical image of an exoplanet, Fomalhaut b, which apparently is mostly solid (Kalas *et al.*, 2008).

A few of the discovered exoplanets are solid. However, according to the law of large numbers, many other telluric exoplanets are expected soon to be discovered from the Earth by techniques such as coronagraphic imaging (*e.g.* Boccaletti *et al.*, 2005) and radial velocity (“gravitational microlensing”), or by the on-going European mission COROT or NASA’s Kepler (Sanderson, 2007). *In situ* investigation or study of exoplanets is not feasible due their remoteness, and no inference about the occurrence of soils is obviously possible at present.

Role of pedology in planetary science

Our nearest planetary neighbours certainly possess complex physically and chemically altered soil mantles. None of these bodies, however, bear (or bore) more resemblance to Earth than does Mars. The periodic return of Martian soil chemical data since the Viking missions of the

1970’s has subsequently sent planetary geologists to acid mine drainages, volcanic mountain tops bathed in acid fog, or to the lab to find Earth analogues that could aid in the interpretation of the Martian data. However, it is the widespread soils of Earth that may hold much insight into the Martian soil history.

Maybe no area on Earth is more relevant to many recent Mars discoveries than the Atacama Desert of northern Chile. The Atacama landscape has endured an exceptionally long and protracted period of aridity (Hartley & Chong, 2002), allowing soils there to achieve ages ($>10^7$ y) impossible on other parts of our otherwise dynamic planet. The soils here lie within a nearly abiotic zone due aridity, providing insights into the protracted arid and presently lifeless Martian history. Soil formation in humid climates is characterised by atmospheric inputs of water and solutes, in-soil biogeochemical weathering processes, and downward leaching of solutes into the underlying vadose zone and groundwater. Soils in hyperarid regions like Chile retain virtually all incoming atmospheric dust and salt, and the infrequent rainfall events that occur over millions of years chemically separate these salts vertically based on solubility (Ewing *et al.*, 2006). This vertical distribution – sulphates over chlorides, nitrates, and perchlorates (Ericksen, 1981) – combined with associated isotopic fractionation during transport (Ewing *et al.*, 2008) – provide clear and unambiguous signals to the long-term net direction of liquid water movement. In addition, oxygenated anions such as nitrate, sulphate, and perchlorate, which are produced by atmospheric reactions with ozone, contain unmistakable isotopic markers of their atmospheric origins. This “mass independent” O isotope signal, well documented in salts found in the hyperarid soils of the Dry Valleys of Antarctica (Bao *et al.*, 2000), has also been detected in sulphates derived from Martian meteorites (Farquhar & Thiemens, 2000), clearly indicating that some similarity in sulphate production mechanisms and transport through the soil zone has operated on both planets. Although far less studied, the Atacama Desert and the Andean Altiplano contain some of the world’s largest sulphate and chloride-rich salt deposits, and the stratigraphy, near surface geochemical profiles, and surface geomorphology of these deposits are untapped sources of information for interpreting the post depositional weathering of salt-rich sedimentary deposits that are now widely recognised on Mars.

If the arid soil beneath our feet provides insight into interpreting available Martian surficial geochemistry, it provides even more important guidance into developing novel hypotheses and new methods for sampling the Martian surface on future missions. First, the direction of water movement through the Martian soils remains, at this stage, a somewhat contentious point of debate (Amundson *et al.*, 2008). Isotopic analyses of salts profiles could address this question as well as provide new insights into the origin of salts. In addition, pH should be more frequently measured on future missions. While jarosite is a mineral that clearly requires acidic formation conditions, the presence or absence of many secondary minerals is at best only weakly suggestive of pH. Geochemical profiles provide a key for interpreting soil history (Ewing *et al.*, 2006; Quade *et al.*, 2007). On Mars, only the fortuitous encounter of the rover Opportunity with Endurance Crater at Meridiani Planum has provided a depth profile of sufficient magnitude required for unambiguous soil geochemical interpretation. Pedological protocols and standardised guidelines for surveying soil are inexpensive in comparison to the costs of a planetary mission. The resulting information could render an unexpected and amplified insight into extraterrestrial soils.

Conclusions

The long term exposure of rocky planets, moons, and other objects within the Solar System and outside to physical and chemical processes has undoubtedly created an array of soil types grading from those impacted by radiation and meteorites to those of Mars which have experienced, in their past, aqueous alteration and transport. On other planets, lacking liquid water, the pace and nature of pedogenic processes is more removed from those we observe on Earth, but these alterations are indeed consistent with a broad, although scientifically constrained, definition of soil.

The growing data on rocky planets in the Solar System opens a window of opportunity for serious discussions about planetary soil formation processes. In addition, some of the trillions of planets that orbit the hundreds of billion stars in our galaxy more than likely possess soils, though their nature may always elude us given the position of our isolated planet in the universe. The consideration of extraterrestrial soils most importantly offers new opportunities for earth scientists to provide quantitative models, data, and insights relevant to interpreting their composition. On the other hand, the Earth is our most accessible natural laboratory, and the full spectrum of its experimental results is required to understand the history of our planetary neighbours.

While our understanding of extraterrestrial soils is very rudimentary, their diversity may raise the question of how we might classify them, or formally compare them with our Earth-based soils. One approach is to simply use our present soil classification schemes, in which case many extraterrestrial soils would be Regosols in the World Reference Base for Soil Resources (IUSS Working Group WRB, 2006) or Entisols in the US Soil Taxonomy (Soil Survey Staff, 2006). On Mars, it is possible that Gelisols and Aridisols exist. However, applying an Earth-based system to such dissimilar settings is debatable. Another option is to distinguish the (largely) biotic Earth from the abiotic Solar System, and include all non-Earth soils in a new Reference Group or Order, which might be tentatively called Astrosols. While we emphasize that classifying extraterrestrial soils is, at this stage, not an important or pressing pedological concern, the discussion helps focus our attention on environments or conditions on Earth that are intergrades to the conditions of our nearby neighbours, and positions pedology to provide increased scientific input towards future planetary missions and data analysis.

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References

Abe, S. *et 15 al.* 2006. Mass and local topography measurements of Itokawa by Hayabusa. *Science,* **312**, 1344–1347.

Amundson, R. 2006. The State Factor theory of soil formation. In: *Soils. Basic Concepts and Future Challenges* (eds G. Certini & R. Scalenghe), pp. 103–112. Cambridge University Press, Cambridge, UK EU.

Amundson, R. *et 8 al.* 2008. On the in situ aqueous alteration of soils on Mars. *Geochimica et Cosmochimica Acta*, **72**, 3845–3864.

Arnold, R.W. 2006. Concepts of soil. In: *Soils. Basic Concepts and Future Challenges* (eds G. Certini & R. Scalenghe), pp. 1–10. Cambridge University Press, Cambridge, UK EU.

Baker, V.R. 2008. Planetary landscape systems: A limitless frontier. *Earth Surface Processes and Landforms*, **33**, 1341–1353.

Bao, H.M., Campbell, D.A., Bockheim, J.G. & Thiemens, M.H. 2000. Origins of sulphate in Antarctic dry-valley soils as deduced from anomalous O-17 compositions. *Nature*, **407**, 499–502.

Barrón, V., Torrent, J. & Greenwood, J.P. 2006. Transformation of jarosite to hematite in simulated Martian brines. *Earth and Planetary Science Letters*, **251**, 380–385.

Barsukov, V.L., Borunov, S.P., Volkov, V.P., Zolotov, M.Yu., Sidorov, Yu.I. & Khodakovsky, I.L. 1986. Mineral composition of Venus soil at Venera 13, Venera 14 and Vega 2 landing sites: Thermodynamic prediction. *Lunar and Planetary Science,* **17**, 28–29.

Bibring, J-P. *et 12 al.* 2005. Mars surface diversity as revealed by the OMEGA/Mars Express observations. *Science*, **307**, 1576–1581.

Bibring, J-P. *et 10 al.* 2006. Global mineralogical and aqueous Mars history derived from OMEGA/Mars Express data. *Science*, **312**, 400–404.

Boccaletti, A., Baudoz, P., Baudrand, J., Reess, J.M. & Rouan, D. 2005. Imaging exoplanets with the coronagraph of JWST/MIRI. *Advances in Space Research*, **36**, 1099–1106.

Britt, D.T., Yeomans, D., Housen, K. & Consolmagno, G. 2002. Asteroid density, porosity, and structure. In: *Asteroids III* (eds W.F.Jr., Bottke, A. Cellino, P. Paolicchi & R.P. Binzel), pp. 485–500. University of Arizona Press, Tucson, AZ USA.

Burns, R.G. 1987. Ferric sulfates on Mars. *Journal of Geophysical Research*, **92**, E570–E574.

Bus, S.J. & Binzel, R.P. 2002. Phase II of the small main-belt asteroid spectroscopy survey: A feature-based taxonomy. *Icarus*, **158**, 146–177.

Castillo-Rogez, J.C., Matson, D.L., Sotin, C., Johnson, T.V., Lunine, J.I. & Thomas, P.C. 2007. Iapetus' geophysics: Rotation rate, shape, and equatorial ridge. *Icarus*, **190**, 179–202.

Chapman, C.R., Morrison, D. & Zellner, B. 1975. Surface properties of asteroids: A synthesis of polarimetry, radiometry, and spectrophotometry. *Icarus*, **25**, 104–130.

Cherkasov, I.I. & Shvarev, V.V. 1978. Current concepts about soils of Mercury, Venus, and Mars. *Soil Mechanics and Foundation Engineering*, **15**, 199–204.

Chevrier, V., Poulet, P. & Bibring, J-P. 2007. Early geochemical environment of Mars as determined from thermodynamics of phyllosilicates. *Nature*, **448**, 60–63.

Cole, G.H.A. 2001. Exoplanets. *Astronomy & Geophysics*, **42**, 1.13–1.18.

Demyk, K., Carrez, Ph., Leroux, H., Cordier, P., Jones, A.P., Borg, J., Quirico, E., Raynal, P.I. & d'Hendecourt, L. 2001. Structural and chemical alteration of crystalline olivine under low energy He^+ irradiation. *Astronomy & Astrophysics*, **368**, L38-L41.

Douté, S., Lopes, R., Kamp, L.W., Carlson, R., Schmitt, B. & the Galileo NIMS Team. 2004. Geology and activity around volcanoes on Io from the analysis of NIMS spectral images. *Icarus*, **169**, 175–196.

Duke, M.B. & Nagle, J.S. 1975. Stratification in the lunar regolith – A preliminary view. *Earth, Moon, and Planets*, **13**, 143–158.

Dukes, C.A., Baragiola, R.A. & McFadden, L.A. 1999. Surface modification of olivine by H^+ and He^+ bombardment. *Journal of Geophysical Research-Planets*, **104**, 1865–1872.

Elachi, C. *et 34 al.* 2005. First views of the surface of Titan from the Cassini RADAR. *Science*, **308**, 970–974.

Ericksen, G.E. 1981. Geology and origin of the Chilean nitrate deposits. USGS Professional Paper 1188.

Ewing, S.A., Sutter, B., Owen, J., Nishiizumi, K., Sharp, W., Cliff, S.S., Perry, K., Dietrich, W., McKay, C.P. & Amundson, R. 2006. A threshold in soil formation at Earth's arid-hyperarid transition. *Geochimica et Cosmochimica Acta*, **70**, 5293–5322.

Ewing, S.A., Wang, W., DePaolo, D.J., Michalski, G., Kendall, C., Stewart, B.W., Thiemens, M. & Amundson, R. 2008. Non-biological fractionation of stable Ca isotopes in soils of the Atacama Desert, Chile. *Geochimica et Cosmochimica Acta*, **72**, 1096–1110.

Farquhar, J. & Thiemens, M.H. 2000. Oxygen cycle of the Martian atmosphere-regolith system: Delta O-17 of secondary phases in Nakhla and Lafayette. *Journal of Geophysical Research-Planets*, **105**, 11991–11997.

Fujiwara, A. *et 21 al.* 2006. The rubble-pile asteroid Itokawa as observed by Hayabusa. *Science*, **312**, 1330–1334.

Fulchignoni, M. *et 42 al.* 2005. In situ measurements of the physical characteristics of Titan’s environment. *Nature*, **438**, 785–791.

Grant, J.A. *et 23 al.* 2004. Surficial deposits at Gusev crater along Spirit rover traverses. *Science*, **305**, 807–810.

Greeley, R. *et 26 al.* 2004. Wind-related processes detected by the Spirit rover at Gusev Crater, Mars. *Science*, **305**, 810–813.

Hapke, B., Cassidy, W. & Wells, E. 1975. Effects of vapour-phase deposition processes on the optical, chemical and magnetic properties of the lunar regolith. *Earth, Moon, and Planets*, **13**, 339–353.

Hapke, B. 2001. Space weathering from Mercury to the asteroid belt. *Journal of Geophysical Research*, **106**, 10039–10073.

Hartley, A.J. & Chong, G. 2002. Late Pliocene age for the Atacama Desert: Implications for the desertification of western South America. *Geology,* **30**, 43–46.

Herkenhoff, K.E. *et 22 al.* 2004. Textures of the soils and rocks at Gusev Crater from Spirit's microscopic imager. *Science,* **305**, 824–826.

Horneck, G. 2008. The microbial case for Mars and its implications for human expeditions to Mars. *Acta Astronautica*, **63**, 1015–1024.

Ishii, H.A., Bradley, J.P., Dai, Z.R., Chi, M., Kearsley, A.T., Burchell, M.J., Browning, N.D. & Molster, F. 2008. Comparison of comet 81P/Wild 2 dust with interplanetary dust from comets. *Science*, **319**, 447–450.

IUSS Working Group WRB. 2006. *World reference base for soil resources 2006, 2nd edition*. World Soil Resources Report 103, FAO, Rome, IT EU.

Kalas, P., Graham, J.R., Chiang, E., Fitzgerald, M.P., Clampin, M., Kite, E.S., Stapelfeldt, K., Marois, C. & Krist, J. 2008. Optical images of an exosolar planet 25 light-years from Earth. *Science*, **322**, 1345–1348.

Keller, L.P. & McKay, D.S. 1997. The nature and origin of rims on lunar soil grains. *Geochimica et Cosmochimica Acta*, **61**, 2331–2341.

Kerr, R.A. 2004. Rainbow of Martian minerals paints picture of degradation. *Science,* **305,** 770–771.

Krasinsky, G.A., Pitjeva, E.V., Vasilyev, M.V. & Yagudina, E.I. 2002. Hidden mass in the asteroid belt. *Icarus*, **158,** 98–105.

Loeffler, M.J., Baragiola, R.A. & Murayama, M. 2008. Laboratory simulations of redeposition of impact ejecta on mineral surfaces. *Icarus*, **196,** 285–292.

Lopes, R.M.C. & Williams, D.A. 2005. Io after Galileo. *Reports on Progress in Physics*, **68**, 303–340.

Lorenz, R.D. *et 39 al.* 2006. The sand seas of Titan: Cassini RADAR observations of longitudinal dunes. *Science*, **312,** 724–727.

Lortet, M-C., Borde, S. & Ochsenbein, F. 1994, The second reference dictionary of the nomenclature of celestial objects. *Astronomy Astrophysics Supplement Series*, **107,** 193–218.

Markevitz, D. 1997. Soil without life? *Nature*, **389,** 435.

Matson, D.L., Johnson, T.V. & Veeder, G.J. 1977. *Soil maturity and planetary regoliths: The Moon, Mercury, and the asteroids*. pp. 1001–1011. Proceedings of the 8th Lunar Science Conference, Houston, TX USA.

Mautner, M.N. 1997. Biological potential of extraterrestrial materials – l. Nutrients in carbonaceous meteorites, and effects on biological growth. *Planetary Space Science*, **45,** 653–664.

McClintock, W.E. *et 12 al.* 2008. Spectroscopic observations of Mercury's surface reflectance during MESSENGER's first Mercury flyby. *Science*, **321**, 62–65.

McCord, T. & Clark, R.N. 1979. The Mercury soil – Presence of Fe^{2+}. *Journal of Geophysical Research*, **84**, 7664–7668.

Melosh, HJ. 2008. Message from Mercury. *Nature*, **452**, 820–821.

Mitri, G., Showman, A.P., Lunine, J.I. & Lorenz, R.D. 2007. Hydrocarbon lakes on Titan. *Icarus*, **186**, 385–394.

Mitrofanov, I. *et 11 al.* 2002. Maps of subsurface hydrogen from the high energy neutron detector, Mars Odyssey. *Science*, **297**, 78–81.

Morris, R. 1980. Origin and size distribution of metallic iron particles in the lunar regolith. *Geochimica et Cosmochimica Acta*, **14 (Supplement)**, 1697–1712.

Morris, R.V. *et 16 al.* 2004. Mineralogy at Gusev crater from the Mössbauer spectrometer on the Spirit rover. *Science*, **305**, 833–836.

NASA. 2009a. *Living on the moon*. URL www.nasa.gov/exploration/ [Accessed on 2009, March 28th]

NASA. 2009b. *Solar system exploration*. URL solarsystem.nasa.gov [Accessed on 2009, March 28th]

NASA. 2009c. *World Book @ NASA*. URL www.nasa.gov/worldbook/ [Accessed on 2009, February 24th]

Navarro-González, R. *et 11 al.* 2008. Mars-like soils in the Atacama Desert, Chile, and the dry limit of microbial life. *Science*, **302**, 1018–1021.

Okada, T., Shirai, K., Yamamoto, Y., Arai, T., Ogawa, K., Hosono, K., Kato, M. 2006. X-ray Fluorescence spectrometry of asteroid Itokawa by Hayabusa. *Science*, **312**, 1338–1341.

OSIRIS Team. 2005. A large dust/ice ratio in the nucleus of comet 9P/Tempel 1. *Nature*, **437**, 987–990.

Porco, C.C. *et 34 al.* 2005. Cassini imaging science: Initial results on Phoebe and Iapetus. *Science*, **307**, 1237–1242.

Poulet, F., Bibring, J-P., Mustard, J.F., Gendrin, A., Mangold, N., Langevin, Y., Arvidson, R.E., Gondet, B. & Gomez, C. 2005. Phyllosilicates on Mars and implications for early martian climate. *Nature*, **438,** 623–627.

Quade, J., Rech, J.A., Latorre, C., Betancourt, J.L., Gleeson, E. & Kalin, M.T.K. 2007. Soils at the hyperarid margin: The isotopic composition of soil carbonate from the Atacama Desert, Northern Chile. *Geochimica et Cosmochimica Acta*, **71**, 3772–3795.

Raulin, F. & Owen, T. 2002. Organic chemistry and exobiology on Titan. *Space Science Review*, **104,** 379–395.

Richter, D.D. & Markevitz, D. 1995. How deep is soil? *Bioscience*, **45**, 600–609.

Robinson, M.S. *et 12 al.* 2008. Reflectance and color variations on Mercury: Regolith processes and compositional heterogeneity. *Science*, **321**, 66–69.

Sanderson, K. 2007. Alien Earth. *Nature*, **445,** 10–11.

Sasaki, S. & Kurahashi, E. 2004. Space weathering on Mercury. *Advances in Space Research*, **33**, 2152–2155.

Schulze-Makuch, D., Irwin, L.N. & Guan, H. 2002. Search parameters for the remote detection of extraterrestrial life. *Planetary Space Science*, **50**, 675–683.

Schulze-Makuch, D. & Irwin, L.N. 2006. The prospect of alien life in exotic forms on other worlds. *Naturwissenschaften*, **93,** 155–172.

Simonelli, D.P., Veverka, J. & McEwen, A.S. 1997. Io: Galileo evidence for major variations in regolith properties. *Geophysical Research Letters*, **24**, 2475–2478.

Soderblom, L.A. *et 43 al.* 2004. Soils of Eagle crater and Meridiani Planum at the Opportunity rover landing site. *Science*, **306**, 1723–1726.

Soil Survey Staff. 2006. *Keys to Soil Taxonomy, 10th edition*. United States Department of Agriculture, Natural Resources Conservation Service. U.S. Government Printing Office, Washington, DC USA.

Solomon, S.C. *et 10 al.* 2008. Return to Mercury: A global perspective on MESSENGER's first Mercury flyby. *Science*, **321**, 59–62.

Squyres, S.W. *et 14 al.* 2008. Detection of silica-rich deposits on Mars. Science, **320**, 1063–1067.

Starukhina, L., Shkuratov, Y. & Skorik, S. 1999. Spread of condensation products in a regolith-like medium: Estimates and laboratory modeling. *Solar System Research*, **33**, 212–215.

Stern, S.A. 1999. The Lunar atmosphere: History, status, current problems, and context. *Reviews of Geophysics*, **37**, 453–491.

Sunshine, J.M. *et 21 al.* 2006. Exposed water ice deposits on the surface of comet 9P/Tempel 1. *Science*, **311**, 1453–1455.

Surkov, Yu.A., Barsukov, V.L., Moskalyeva, L.P., Kharyukova, V.P. & Kemurdzhian, A.L. 1984. New data on the composition, structure, and properties of Venus rock obtained by Venera 13 and Venera 14. *Journal of Geophysical Research*, **89 (Supplement)**, B393–B402.

Swain, M.R., Vasisht, G. & Tinetti, G. 2008. The presence of methane in the atmosphere of an extrasolar planet. *Nature*, **452**, 329–331.

Taylor, L.A. *et 10 al.* 2002. Martian meteorite Dhofar 019: A new shergottite. *Meteoritics & Planetary Science*, **37**, 1107–1128.

Thomas, P.C. *et 12 al.* 2007. Shapes of the saturnian icy satellites and their significance. *Icarus*, **191**, 371–380.

Treiman, A.H., Barrett, R.A. & Gooding, J.L. 1993. Preterrestrial aqueous alteration of the Lafayette (SNC) meteorite. *Meteoritics*, **28**, 86–97.

Trombka, J.I. *et 24 al.* 2000. The elemental composition of asteroid 433 Eros: Results of the NEAR-Shoemaker x-ray spectrometer. *Science*, **289**, 2101–2105.

Udry, S. *et 10 al*. 2007. The HARPS search for southern extra-solar planets^{H} XI. Super-Earths (5 & 8 $M_{\varnothing}$) in a 3-planet system. *Astronomy & Astrophysics*, **469**, L43-L47.

Ugolini, F.C. & Anderson, D.M. 1973. Ionic migration and weathering in frozen Antarctic soils. *Soil Science*, **115**, 461–470.

Ugolini, F.C. & Bockheim, J.G. 2008. Antarctic soils and soil formation in a changing environment: A review. *Geoderma*, **144**, 1–8.

Ugolini, F.C. & Edmonds, R.L. 1983. Soil biology. In: *Pedogenesis and Soil Taxonomy. I. Concepts and Interactions* (eds L.P. Wilding, N.E. Smeck, G.F. Small), pp. 193–231. Elsevier Publishers, Amsterdam, NL EU.

van Boekel, R. *et 22 al.* 2004. The building blocks of planets within the “terrestrial” region of protoplanetary disks. *Nature*, **432**, 479–482.

Vaniman, D.T., Bish, D.L., Chipera, S.J., Fialips, C.I., Carey, J.W. & Feldman, W.C. 2004. Magnesium sulphate salts and the history of water on Mars. *Nature*, **431**, 663–665.

Veverka, J. & Burns, J.A. 1980. The moons of Mars. *Annual Review of Earth and Planetary Sciences*, **8**, 527–558.

Weber, P. & Greenberg, J.M. 1985. Can spores survive in interstellar space? *Nature*, **316**, 403–407.

Wilson, P.D. & Sagan, C. 1996. Spectrophotometry and organic matter on Iapetus. 2. Models of interhemispheric asymmetry. *Icarus*, **122**, 92–106.

Yen, A.S., Kim, S.S., Hecht, M.H., Frant, M.S. & Murray, B. 2000. Evidence that the reactivity of the martian soil is due to superoxide ions. *Science*, **289**, 1909–1912.

FIGURE CAPTIONS

Figure 1. The footprints left by the astronauts in the powdery Sea of Tranquillity are more permanent than most solid structures on Earth. Barring a chance meteorite impact, these impressions in the lunar soil will probably last for millions of years. [Courtesy of NASA]

Figure 2. Terrestrial environments formed under virtually abiotical conditions: a) the Wright Valley, McMurdo Dry Valleys System, Antarctica [courtesy of FC Ugolini], and b) the Atacama Desert, northern Chile [photo by R Amundson].

Figure 3. Image of Mercury taken by the NASA's spacecraft Messenger, showing a variety of surface textures, including smooth plains at the centre, numerous impact craters and rough material that appears to have been ejected from the large crater to the lower right. [Courtesy of NASA]

Figure 4. Venus' surface pictured by the Venera 13 lander (URL nssdc.gsfc.nasa.gov/planetary/venera.html). [Courtesy of NASA]

Figure 5. Astronaut Alan Bean drives a soil core tube into the Lunar soil surface during the Apollo 12 mission (photo #AS12-49-7286 from the Project Apollo Archive, URL www.apolloarchive.com/apollo_gallery.html). [Courtesy of NASA History Office]

Figure 6. Mars: a) A portion of the "Presidential Panorama" of the Pathfinder landing site, with light-toned sulphate rich outcrops in lower left and top, left/center (URL mars.jpl.nasa.gov/MPF/ames/ames-pres.html); b) Patterned ground observed at the Phoenix mission landing site. [Courtesy of NASA]

Figure 7. Martian soil profiles (a) and (b) exposed by NASA's Phoenix, and the Martian soil lab (c) equipped with microscopy, electrochemistry, and gas analyzer. [Courtesy of NASA/JPL-Caltech/University of Arizona/Texas A&M University/Max Planck Institute]

Figure 8. The Mars' moon Phobos pictured on 2004 by the Mars Express spacecraft, from some hundreds kilometers. [Courtesy of ESA/DLR/FU Berlin G. Neukum]

Figure 9. 25143 Itokawa. Lateral image, 0.5 by 0.3 by 0.2 kilometres. The asteroid is composed of discrete rock fragments and fine earth held together by gravity. [Courtesy of JAXA]

Figure 10. Saturn's moons Titan and Iapetus: a) Rivers on Titan's landscape; b) The white frozen Iapetus spotted with carbonaceous dark areas pictured by the NASA Cassini orbiter in 2007. [Courtesy of ESA/NASA/JPL/Space Science Institute/University of Arizona]

Appendix. Main characteristics of telluric bodies in space.

		Mercury	Venus	Earth	Moon	Mars	Phobos[1]	Deimos[1]	Asteroids[2]	Io[3]	Titan[4]	Iapetus[4]	[l]Comets[5]	Exoplanets[6]
Physical characteristics	[b, t]diameter /km	~4900	~12100	~12700	~1700	~6800	~20	~10	1-1000	~3600	5600	~1600		
	[b]md[10] /kg m^{-3}	5.4	5.2	5.5	3.3	3.9	1.9	2.2	[g]1.0-4.9	3.5	1.9	[p]1.1	?	?
Detection[#]		⚟	☑	☑	☑	☑	⚟	⚟	[h]☑, ☐, ⚟	☐	☑	☑	⚟	⚟
Atmosphere	[b]mst/[b]MST[7] /K	100/740	750	180/310	40/400	150/310	230	230	200?	[u]80/140	90	100/130	?	?
	thermal excursion[8]	↓↑/↓↑	↓↑/↓↑	↓↑/↓↑	=/↓↑	↓↑/↓↑	=/↓↑	=/↓↑	?/?	=/↓↑	=/↓↑	=/↓↑		?/?
	albedo	0.10	0.59	0.39	0.12	0.15	0.07	0.07	0.10-0.22	0.62	0.21	0.3-0.6	?	?
	[s]pressure /kPa	10^{-15}	~9100	~100	10^{-13}	~1	$<10^{-15}$?	$<10^{-15}$?		[u]10^{-9}	[t]150	?		
	[a, c]main components /%	$He/H_2/O_2$	$CO_2/N_2/O_2$	$N_2/O_2/Ar$	[e]CH_4 tr.	$CO_2/N_2/Ar$	?	?	?	SO_2	$N_2/Ar/CH_4$	?	?	?, [r]CH_4
Hydrosphere[‡]		🌧	🌧	💧, ✱, 🌧	-	✱, 🌧	✱	(💧)	-	[i]✱	(💧)	[q](✱)	[n](✱)	(💧)
	[a]liquid solvent	none	H_2O, H_2SO_4	H_2O	none	H_2O	H_2O?	H_2O?	none	H_2S	[j]CH_4	[q]NH_3, ?	none	?
Lithosphere	[a, c]lithology[9]	b, a	b, am	b, g, am	b, a	b, am	cc	cc	[f]cc, s, m	s	?	[b]cc (?)	[m]s (?)	?
Pedosphere[†]		✦, (✶,★)	✦, (★)	✹	✶, (✹)	✶, (✹)	✦	✦	★, ✦	✦	*[✦]*	*[✦]*	*[✦]*	✦
[b]Biosphere[*]		☹	😐	☺	☹	😐	☹	☹	☹	☹	[k]😐	😐?	☹	😐?

[1] Satellites of Mars.

[2] 376,537 are registered as minor planets, their global mass is estimated to be about 3.0-3.6×10^{21} kg[d]. Originally, only three types of asteroids were classified: C-type carbonaceous, (75% of known asteroids), S-type silicaceous, (17%), L-type metallic, (8%). More recently 26 types have been defined[o].

[3] Satellite of Jupiter.

[4] Satellite of Saturn.

[5] According to the International Astronomical Union comets are designated by the year of their discovery followed by a letter indicating the half-month of the discovery and a number indicating the order of discovery.

[6] Planets beyond the Solar System.

[#] Detection: landing and collecting samples ☑, research with space probes, not collecting samples ☐, remote sensing only ⚟

[‡] Hydrosphere: liquid water 🌢, solid water ✱, vapor ⛆, presence of liquid water (🌢) or ice (✱) inferred.

[†] Pedosphere: active pedogenesis and horizonation ✹, physical reworking only ✶, chemical weathering ★, pedogenesis plausible ✦, pedogenesis improbable *[✦]*.

[*] Biosphere: active ☺, not detected but possible 😐, not likely to be present based on current knowledge ☹

[7] Atmosphere characteristics: Maximum Surface Temperature MST, minimum surface temperature mst.

[8] Thermal excursion (difference between the highest temperature and the lowest one in a given lapse of time): ↓↑ occurrence, = absence; the symbol on the left of the slash indicates diurnal thermal excursion, that on the right seasonal thermal excursion.

[9] Lithosphere: basalts b, granits g, anorthosite a, altered materials am, carbonaceous chondrites cc, silicates s, metallic m.

[10] Lithosphere: mean density md.

Sources: a) Schulze-Makuch *et al.* (2002); b) NASA (2009b); c) Wikipedia; d) Krasinsky *et al.* (2002); e) Stern (1999); f) Chapman *et al.* (1975); g) Britt *et al.* (2002); h) Okada *et al.* (2006) i) Douté *et al.* (2004); j) Mitri *et al.* (2007); k) Raulin & Owen (2002); l) Lortet *et al.* (1994); m) van Boekel *et al.* (2004); n) Sunshine *et al.* (2006); o) Bus & Binzel (2002); p) Thomas *et al.* (2007); q) Castillo-Rogez *et al.* (2007); r) Swain *et al.*, 2008, s) NASA (2009c); t) Fulchignoni *et al*. (2005); u) Lopes & Williams (2005).