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Creating Habitable Zones, at all Scales, from Planets to Mud Micro-Habitats, on Earth and on Mars

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Abstract The factors that create a habitable planet are considered at all scales, from planetary inventories to micro-habitats in soft sediments and intangibles such as habitat linkage. The possibility of habitability first comes about during accretion, as a product of the processes of impact and volatile inventory history. To create habitability water is essential, not only for life but to aid the continual tectonic reworking and erosion that supply key redox contrasts and biochemical substrates to sustain habitability. Mud or soft sediment may be a biochemical prerequisite, to provide accessible substrate and protection. Once life begins, the habitat is widened by the activity of life, both by its management of the greenhouse and

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by partitioning reductants (e.g. dead organic matter) and oxidants (including waste products). Potential Martian habitats are discussed: by comparison with Earth there are many potential environmental settings on Mars in which life may once have occurred, or may even continue to exist. The long-term evolution of habitability in the Solar System is considered.

Keywords Mars · Water · Fluvial erosion · Habitable surfaces · Rock cycle

1 Introduction

This chapter follows the creation of habitability on all scales, on Earth and on Mars. It examines the creation of habitability on a planetary scale, then the formation of a suitable surface environment, then focuses down to the formation of local zones, such as muds, where habitation could occur. Habitability is not just macro-suitability, having a planet in the right place next to the right star. It is also micro-scale: the availability of cell-sized, cell-friendly habitats. Moreover, habitability must be sustained over billions of years—fortunately life has self-sustaining character.

The focus is on the one planet we know to be habitable: the Earth. Mars is also discussed towards the end of the chapter. The implications range more widely: to be habitable, extrasolar planets will need to be suitable both on the grand planetary scale, and on the micro scale.

A habitable planet must be able to sustain life over eons. The fundamental environmental requirements for habitability include providing favorable conditions for the assembly of organic molecules and energy sources to sustain metabolism. Less obvious is that far more than offering a single specific environment (which will obviously be short-lived), habitability needs diversity. To continue, life needs an ongoing suite of environments that communicate through the exchange of materials and which are sustained in time.

A habitable planet must "live" geologically. It must renew its surface. It has to be active volcanically and tectonically. Habitable environments must provide a highly flexible exchange medium for chemical and biochemical components. Although it is conceivable that other solvents could carry out this role (e.g. ammonia), the obvious (and possibly unique) medium for exchange is water (Brack 2002). If so, the planetary environment must sustain liquid water. Water's physical prerequisites constrain the abundance and distribution of habitable zones, managing every aspect from hydrothermal systems to UV control.

During the earliest Hadean, the primary construction of the planets was accomplished. The sites for the potential houses for life were marked out, each with its distinct power service from the Sun's light. The foundations accreted, planetesimal by planetesimal; the bodies were put up by much bashing and crashing. The plaster was thrown onto the walls. Then in the mid-Hadean finishing-work on the various rooms for life was begun. What was rough and uninhabitable slowly took on the appearance of a living room, where life could make itself comfortable. Life appeared and claimed its habitat—at least on Earth, perhaps on Mars; just possibly on Venus. And once it had arrived, life began changing the room immediately: not only did it fill the room with furniture, but it began to replaster the walls and eventually even to remodel the foundations themselves.

In this chapter we follow the story from the end of the mid- to late-Hadean, around 4.2 to 4 Ga ago, when the main work of accretion was completed but some bombardment was still continuing, to the beginnings of what we recognise as a 'modern' oxic habitat, $\sim$2.3 Ga ago. The central focus of the account is water and its interactions with the planet and with life. The first part of the chapter discusses the degassing of the oceans and the interaction between water and rock. Then the water-rich habitable environments are explored, and the sources of the chemical disequilibrium that supported early life. The early record of life on Earth is explored: what signatures has early life left in the rock record and in our own genes in modern life? What impact did life have in remodeling its home? Mars is explored—are there such habitats there. Finally, the issue of sustained habitability is discussed.

The Zimbabwean geologist A.M. Macgregor (1927), pointed out that the modern atmosphere of the Earth is a biological construction. What came before presumably had far more carbon dioxide; life remade the air. Macgregor also pointed out that in changing the air, life also reshaped the controls on greenhouse temperature. It risked changing life-giving water into hard rock, setting off glaciation. In general, life shapes, changes and sustains its own conditions.

2 The Planetary Scale: Creating Habitability

For long-term habitability by carbon-based life, a planet needs both:

1) Liquid water to provide a medium of chemical exchange (Brack 2002), and
2) Sustained thermodynamic disequilibrium between chemical species in close spatial proximity. The latter is most effectively done by utilizing solar photons, which are grossly out of equilibrium with conditions on the Earth.

Water is the most abundant condensable material in the cosmos. It is arguably the likeliest chemical compound to be delivered to a new planetary surface. Here it is taken as granted that the early Earth and many other planetary bodies in the solar system had significant water inventories at the end of accretion.

The problem for the biologist is that at least some of this water must be on the surface of the planet, not dissolved in its interior. Indeed, for many years the lifespan of oceans remained an unsolved problem for geologists over time, why does the water not simply sink into the ground and vanish? That is exactly what it has mostly done on Mars. Only with the discovery of plate tectonics did the answer come for Earth—because the Earth's interior is hot, hydrated oceanic crust does indeed sink in, but the water is recycled by partial melting in subduction zones.

Water exists as a liquid at pressures above 6.1 mbar and temperatures above 273 K. On planetary surfaces temperatures are determined by the solar irradiance, the planet's albedo (reflectivity), and the atmosphere's greenhouse effect. Both the albedo and the greenhouse effect depend in turn on the response of water to surface temperature. The albedo is sensitive to ice, snow and clouds, while the greenhouse effect is dominated by the opacity of water vapor to thermal radiation. The chief independent variables are the irradiance and the abundance of other greenhouse gases. The latter can be something of a wild card, but the irradiance is a simple well-defined function of the stellar luminosity (L) and distance (D) from the star: L/D^2. Thus, phase conditions suitable for liquid water depend on the distance from the central star. Conservative estimates for the innermost and outermost distances where liquid water is stable in our solar system are 0.95 and 1.37 AU, respectively (Kasting et al. 1993). This range of distances has been called the "habitable zone".

It is noteworthy that this habitable zone moves outwards from the Sun over time. This is because the Sun becomes brighter as it ages. The Sun is now about 30% brighter than it was 4 Ga (see Fig. 1, Zahnle et al. chapter, this volume) ago. We will discuss the deep past and the deep future in more detail at the end of this chapter.

The inner edge of the habitable zone is hard to finesse short of painting the planet white, but the outer limit is negotiable, provided that there be greenhouse gases enough. Popular

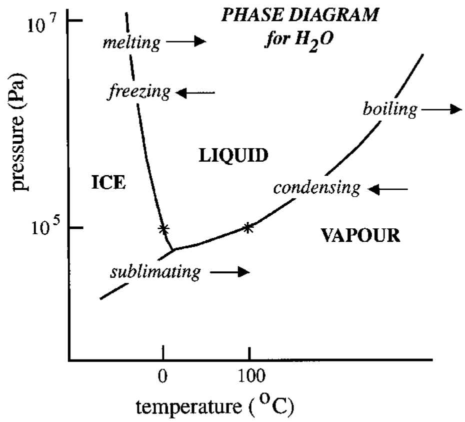


Fig. 1 Phases of H_2O (triple point is at 6.1 mbar and 273 K)

candidates include CO_2, NH_3, and CH_4, but there are many more a technological civilization can manufacture new ones (e.g. C_3F_8) by the dozens. Stevenson (1999) pointed that with enough hydrogen in the atmosphere there may be no effective outer limit at all.

Another option is provided by internal heat sources (such as radioactive or tidal heating). With heat welling up from below planetary crusts can provide protected insulated subsurface conditions that allow a liquid phase of H_2O in planetary bodies even outside the habitable zone of liquid water. Therefore subsurface habitable environments may be a major target in searching for extant and extinct extraterrestrial life, if we could access them, for example by targeting outflow regions. Otherwise, the search must be confined to surficial life and the superficial qualities of life, until deep drilling on Mars or Europa becomes possible. Outcrops with evidence of fossil subsurface life may be found on present planetary surfaces, however, because to some extent impact craters exhume underlying material. Shallow drilling ($<$5 m) in Martian gully regions might also be a possibility to look for organics.

3 Making a Habitable Zone: The Planetary Skin in the Hadean

Chapter 2 has shown how the planet was assembled. By 4.4 Ga ago it had a cool surface with liquid water: the first prerequisite for habitability.

A cool surface is not enough. There must be chemical differences that life can exploit. On the Hadean Earth and early Mars this was provided by the thermodynamic difference between atmosphere/hydrosphere chemical species created by solar photons, entering the top of the atmosphere, and those interacting with magma from the interior. The interior provided air, and below it water, and cycled that water in hydrothermal systems. The exterior setting in the solar system provided energy—photons—that reworked the components of the ocean/atmosphere system, creating chemical species out of equilibrium with the interior.

The temperature of the interior is crucial. On modern Mars, it is low. On early Earth it was high. The potential temperature of the mantle is the temperature of a piece of the mantle if it were suddenly brought unmelted to the surface. Immediately after the Moon-forming event, temperature would have been very high, above 2500 K. By the late Archean the "potential temperature" of the mantle ("potential temperature" is a thermodynamic term defined as the temperature at which magma would rise adiabatically to the surface) had fallen to about 1800 K (Nisbet et al. 1993). Today the potential temperature is about 1600 K.

To form a solid lid on the Earth's surface, over a molten interior, the surface temperature must fall below about 2000 K (Sleep et al. 2001). By the time a first proto-crust formed, the potential temperature was about 2500 K (Sleep et al. 2001). This was probably the potential temperature before 4.4 Ga ago, at a time a few millions of years after the Moon-forming event. Sleep et al. (2001) suggest that this could have been around 3 million years after the global melting event, assuming a massive CO_2 atmosphere did not exist. There would then follow an interval during which the surface heat flow from the interior, at around 70–100 W m^{-2}, would have been sufficient to maintain equable surface conditions around 30°C. For comparison, the modern Earth receives a very roughly comparable heat input from the Sun. However, such an environment would be barely habitable—the crustal rind would be some tens of meters thick, with molten rock below. This would not be mechanically stable, and any newly-born living cells would be soon tipped into the inferno by minor quakes.

3.1 Hadean Surface Temperature: The Cool Early Earth

The oldest material on the Earth is found in the interiors of single crystals of zircon, some of which formed as early as 4.4 Ga ago (Wilde et al. 2001). From the δ^{18}O values of these zircons, some broad generalizations about conditions on the planetary surface conditions can be inferred. Their titanium contents, used as a geothermometer, provide evidence of formation in water-rich granitic magmas (Watson and Harrison 2005). The only unusual aspect of the 4.4 Ga zircons is their age. In all other respects they are closely similar to later (Archean) zircons. From this similarity it is inferred that they incorporated oxygen from water that had been in contact with the planetary surface.

The chain of logic is as follows: 1) δ^{18}O values in zircons are nearly constant for the long time interval 4.4 to 2.6 Ga ago. 2) liquid water oceans must have been present throughout the 2.6 Ga to 3.8 Ga period, to deposit the clastic and chemical sediments that are ubiquitous in the Archean record. 3) Zircons made in the late Archean incorporate water from hydrothermal systems linked to liquid oceans. Ergo 4) the Hadean zircons, that are closely similar to the Archean examples, also incorporate oxygen from liquid oceans. Thus the conclusion is that liquid water existed as early as 4.4 Ga ago.

From this type of reasoning, Valley et al. (2002) inferred a surface temperature of lower than 200°C at 4.4 Ga. This is consistent with the finding of Sleep et al. (2001), that if the entire CO_2 inventory were placed in the air, the temperature would have been around 230°C.

Sleep et al. (2001) showed the transition from massive CO_2 greenhouse to cool surface temperature probably took place rapidly but not instantaneously. Widespread surface conditions hospitable to hyperthermophilic microbial organisms (say around $\sim$100°C) probably persisted for a few million years (range <1 to $\sim$20 million years). Thus for a brief period, probably around 1 million years but possibly as long as 20 million years, it might be possible to maintain surface temperatures around 100°C, suitable for hyperthermopile life. After the close of this interval, the system would cool and the surface temperature would begin to drop, eventually towards glacial conditions. In a late Hadean icehouse, hyperthermophile habitats with temperatures around $\sim$100°C would be tightly restricted to the hydrothermal systems in the vicinity of volcanic activity.

Nevertheless, the 'brief' <1 to $\sim$20 million years of widespread warm oceans is a very long time, especially when thought of in terms of microbial generations. Any line of living organisms born at this early time would be preadapted to take refuge in hydrothermal systems as the world ocean cooled. Such an early hyperthermophile biota would be likely to suffer one or more ocean-boiling meteorite impacts before the end of the Hadean at $\sim$4 Ga. Nevertheless, cells might survive a global heating event if they had refuge in the rock, in hydrothermal systems.

3.2 The Lively Interior: The Contribution from Volcanism

Hadean volcanism would have occurred in an already formed ocean, as demonstrated by the zircon oxygen isotope evidence (Valley et al. 2002; Wilde et al. 2001). After each mid-Hadean impact (a 10 km-size body each 0.1 Ma at that time), massive quantities of basaltic and komatiitic ejecta would have been thrown out and then landed back into the seas. Massive komatiitic volcanism was probably occurring, locally intensified in the aftermath of major impacts. Consequently there would be large-scale interaction of ocean water with hot mafic rock. This hydrothermal interaction would have been much more rapid than today: perhaps such that every million years or less a volume of water equal to the volume of the oceans would pass through hydrothermal systems.

The consequence of this would be pH and geochemical control on the water body. Close to hot vents, hydrothermal fluids would be acidic, but komatiitic volcanism was very widespread and the dominant cooler exiting flow may have been more alkaline. As far as prebiotic conditions are concerned, it is likely there was an array of settings, from highly alkaline around cool hydrothermal systems in ageing ultramafic lavas, to highly acid in proximal settings to active basaltic volcano vents.

Elsewhere in this book (Chap. 1) we argue that, other things equal, the faint young Sun would render surface conditions cool to icy, except immediately after very large-scale impact events. Ice is a good insulator, and ambient geothermal heat would be trapped underneath it. Thus an ocean that was 5 km thick would not be frozen solid. Even today, Antarctica has huge lakes under the ice for this reason: the ambient geotherm. Volcanic vents would add to this ambient heat, and would have provided warm oases, where liquid water might reach the surface. Typical Hadean heat flows would have kept the ice $\sim$100 m thin, and over ridge axes or hotspots the ice would have thin enough ($<$10 m) that sunlight would have played a major role in the ice sheet's heat budget.

Even if the bulk of the planetary surface was ice, some pools would exist around the abundant active volcanoes (especially given the higher potential temperature of the mantle) or as leads between jostling ice masses. The zircon evidence for subduction implies that crustal blocks sank toward the mantle, which suggests that new crust was forming at the surface (as opposed to say a stagnant lid regime). We argue elsewhere in this book that Hadean tectonics would have superficially resembled plate tectonics in having sea-floor spreading and subduction. Hadean mid-ocean ridges may have been komatiitic, with huge far-running lava flows. Hence large pools of liquid water would be expected along the lengths of the mid-ocean ridge systems, perhaps under thick icy cover. Because mid-ocean ridges are linear and connected, there could have been major warm sub-ice lakes.

On very early Venus, D/H evidence implies a warm or hot ocean existed, liquid throughout, perhaps several km deep (Watson et al. 1984). On Mars, conditions would be more like the Hadean Earth, with short-lived hot phases after major impacts, but with a puddle ocean and active tectonics (Sleep 1994). The volcanism would be less vigorous and perhaps somewhat cooler—dominantly mafic rather than ultramafic. Heat flow would have declined faster. Impacts would create major topographic features such as depressions and lakes. Impacts would also cause porosity in subsurface rocks, which would be taken up by hydrothermal fluids. On Mars this porosity would extend several km deep. On Earth, with higher gravity and temperatures, porosity would anneal better.

Around hydrothermal system vents in all three planets, muds would have formed. On early Earth muds were probably dominantly palagonitic, rich in smectite clays (nontronite, saponite, etc.) characteristic of pillow debris. However the presence of old zircons implies that locally at least there were granitoid bodies, and thus it is likely that some muds were felsic-derived clays.

4 Impact Modeling—Tilling the Surface: Processing the Habitability Zone

A young planet experiences major impacts. Each such major impact would have tilled the surface of the planet. Very large impacts destroy extant habitability, though may increase it later. More modest impacts can enhance habitability. Craters exhume material and thus investigation by drilling may not always be warranted: the experiment is already done.

The history of impacts on early Earth is discussed in the chapter by Zahnle et al. in this book. The Moon is the chief source of data. The face of the Moon has been sculpted by uncounted myriads of impact craters, ranging from vast basins thousands of kilometers across to the microscopic. Earth, because it is bigger, was hit by twenty projectiles for every one that hit the Moon, and with a high degree of confidence we can assert that Earth was hit by several projectiles that were much bigger than any that ever hit the Moon. The big ones posed a serious recurrent hazard to life on Earth (Zahnle and Sleep 1997, 2006).

Figure 2 shows a model realization of the aftermath of a very large impact on Earth. The model is based on conservation of energy and the radiative properties of steam atmospheres. It is akin to, but simpler than, the model of Earth's cooling history following the Moon-forming impact described in the chapter by Zahnle et al. (this volume). The key features of the model is that it presumes that a significant fraction of the impact energy of a very large impact is invested in making and spreading rock vapor around the planet. The rock vapor cools by radiating upward to space or downward onto the seas. In a big impact the latter is the bigger term because the rock vapor atmosphere is hotter at the bottom than at the top. As the rock vapor cools it condenses into rock raindrops that fall out and quench in the seas. The energy absorbed by the seas—the thermal radiation and the cooled molten rain—evaporates water. The water vapor builds up in the atmosphere until the energy of the rock vapor is exhausted. Thereafter the steam atmosphere cools radiatively until the excess water

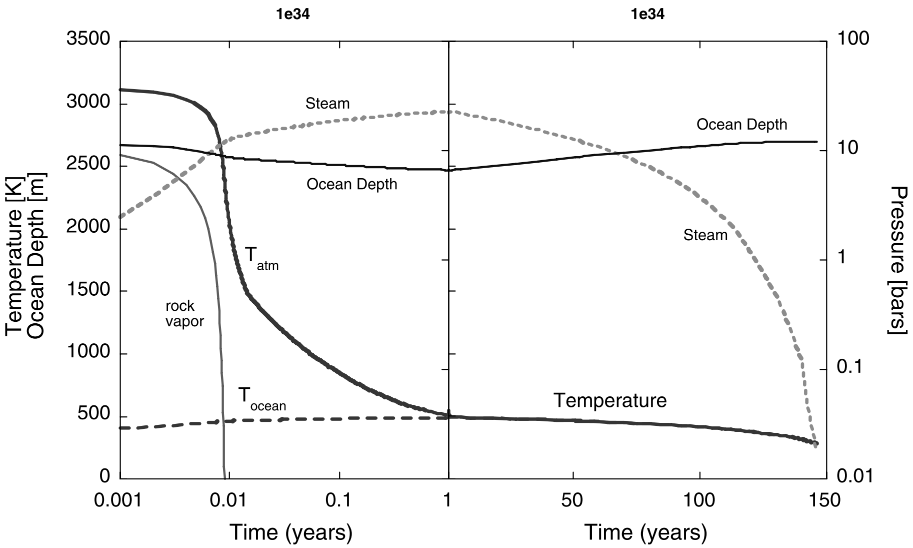


Fig. 2 Aftermath of a S. Pole-Aitken or Hellas scale impact on Earth. The energy released is 10^{27} joules. The figure plots ocean depth, sea surface temperature, atmospheric temperature, and the gas pressures of rock vapor and steam while in the atmosphere. The modern terrestrial atmosphere contains 1 bar of nitrogen and oxygen

vapor has fully rained out. How long this takes is governed by the thermal energy of the atmosphere, the latent heat of condensation of the water vapor, and the radiative physics of the runaway greenhouse effect (see the chapter by Zahnle et al., this volume). More details describing these sorts of models are given by Zahnle and Sleep (1997, 2006) and Sleep and Zahnle (1998).

The impact modelled in Fig. 2 is comparable to the largest impacts recorded in the extant crust of the solar system: those which made the South Pole-Aitken basin on the Moon (2500 km wide, 13 km deep) or the Hellas Planitia (2100 km wide, 8 km deep) basin on Mars. The energy released in these impacts was on the order of 10^{27} joules. The figure plots ocean depth, sea surface temperature, atmospheric temperature, and the amount of steam and rock vapor in the atmosphere as a function of time.

The response occurs on two basic time scales—a period of a few days during which rock vapor is present in the atmosphere, and a longer period of more than 100 years during which the atmosphere contains abundant hot steam. Energy in the rock vapor evaporates a few hundred meters of seawater. The surface waters of the ocean heat to nearly 500 K, but the deep waters can remain cool. Hot fresh rainwater pools at the top of the ocean, so that the ocean is stably stratified both by temperature and by composition.

An impact on this scale would have been disastrous for photosynthetic life: in short, the habitat zone would be eliminated. However, hydrothermally-hosted life in deep water might have survived.

Figure 3 shows the results of a slightly smaller impact, on the scale of that which made the ~1000 km wide lunar basins, such as Mare Imbrium or Orientale. These basins are prominent because they host lunar maria—dark basaltic lavas that flooded the low parts of the Moon and make up the pattern of light and dark known as the "Man in the Moon". The energy released is 2×10^{26} joules. As before, there are two basic time scales – a day or so during which rock vapor is present in the atmosphere, and a longer period of 30 years

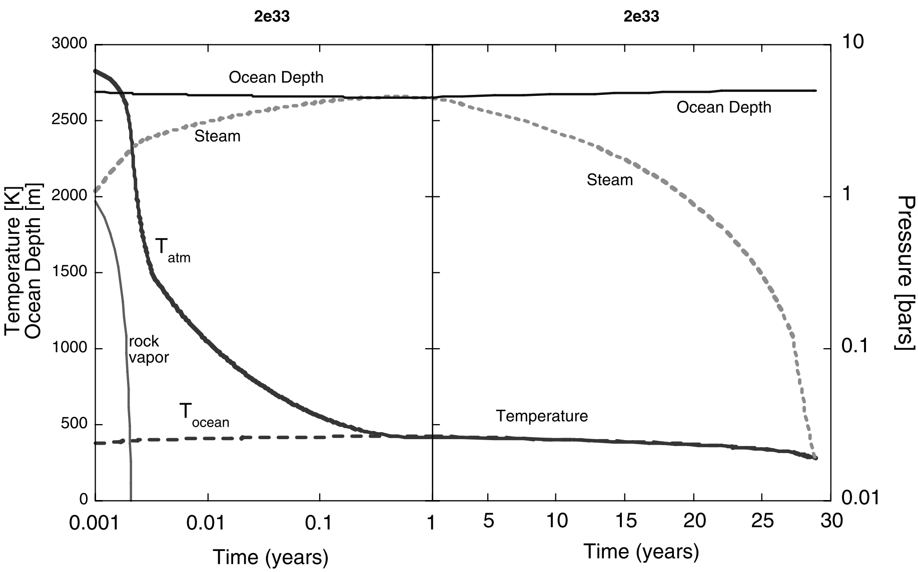


Fig. 3 Aftermath of an Imbrium or Orientale scale impact on Earth. The energy released is 2×10^{26} joules. The figure plots ocean depth, sea surface temperature, atmospheric temperature, and the amount of steam and rock vapor in the atmosphere

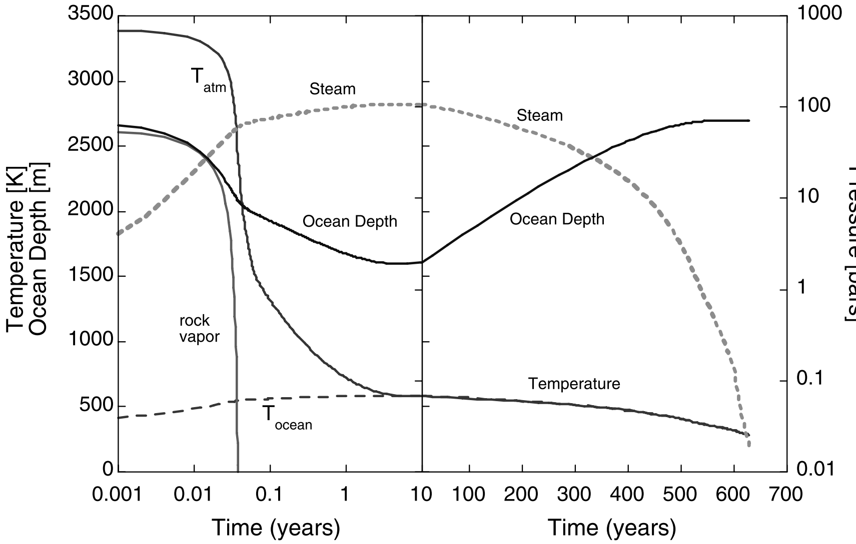


Fig. 4 Aftermath of a hypothetical 5×10^{27} joule impact of on Earth. A small number of impacts on this scale are likely on Earth at the time when the lunar Imbrium and Orientale impact basins formed. The figure plots ocean depth, sea surface temperature, atmospheric temperature, and the amount of steam and rock vapor in the atmosphere

during which the atmosphere contains abundant hot steam. In this case, energy in the rock vapor evaporates only 50 meters of seawater. The surface waters of the ocean heat far less, to nearly 150°C, but the deep waters remain cool. As before hot fresh rain waters pool at the top of the ocean, so that the ocean is stably stratified both by temperature and by composition, though less so than in the case of the larger impact. Even so, contemporary photosynthetic communities may have fared ill.

Figure 4 considers an impact even bigger than the Hellas and the S. Pole/Aitken events. The figure illustrates the aftermath of a hypothetical 5×10^{27} joule impact on Earth. A small number of impacts on this scale (perhaps a half-dozen) are likely on Earth at the time when the lunar Imbrium and Orientale impact basins formed. Aside from the bigger scale of the event, the figure is directly analogous to Figs. 2 and 3. The two basic time scales here are a few weeks for rock vapor and 600 years for an atmosphere of hot steam. There is enough energy in the rock vapor to evaporate more than a kilometer of seawater. The surface waters of the ocean heat to 550 K. Boiled brine sinks into the ocean and contains enough energy to heat the unevaporated sea waters by 50 K. Whether the deep oceans actually do get hot depends on whether the oceans mix. The deep waters are at first cold and dense and therefore stable against convection. However the oceans may mix by other means, for example through shear instabilities if currents are strong, as they might be expected to be. Later, hot fresh rainwater would tend to pool on top of the remnant salty ocean. Again, static stability would inhibit convection, and it is possible albeit not guaranteed that the hot fresh waters would long remain afloat atop the colder denser remnant waters. Ecological consequences would be severe and mesophilic ecosystems may not survive.

5 Adding Components of Habitability to the Planetary Skin: The Impact Heritage on the Surface Inventory

Giant impacts totally rework the planetary body (Sleep et al. 2001). The largest (bodies of hundreds of km diameter) create a magma ocean and a huge steam or molten silicate atmosphere. Were one to have occurred in the past 3.5 Ga (billion years), life would have been eliminated. Large impacts (say 5–50 km diameter) do not terminate microbial life, but can cause mass extinction bottlenecks, even among prokaryotes. Impacts on a smaller scale do not produce any global changes but can provide changes in local environment, which can be favorable for habitability (warm ponds, nutrients, etc.) (Sleep et al. 2001).

The volume of a crater produced by a hypervelocity impact is hundreds of times larger than that of the projectile. Most of that volume is fractured rocks, which fall to surface near the crater. Solid state shocked rocks can contribute to habitability by creation of multiple rock cracks with fresh (i.e. chemically active) surfaces, moderate warming, and possibly by some effects of shock metamorphism (decomposition of volatile bearing minerals, mobility of rock-forming elements, etc.).

The volume of melted rock is from several to tens of times of the volume of the projectile. Impact-induced melting releases incorporated volatile components. Part of the melt is deposited in large craters as sufficiently well mixed and rapidly solidified bodies, which provide long term heating of subsurface crater material stimulating hydrothermal activity. For typical asteroidal impacts (with velocities of 10–25 km/s) silicates are not completely vaporized (Pierazzo and Melosh 2000). Partial vaporization creates a gas phase of volatile elements and leaves refractory elements in the residual melt. The vapor is about equal in mass to the projectile. This vapor forms fireball ejecta with starting temperatures up to 5000–6000 K and pressures up to several thousands of bars. As the vapor plume expands it cools, and condenses.

Gerasimov et al. (1999) have shown that the end products after expansion and quenching at high-temperature appear to be formed in strong disequilibrium, quite different from products expected in the normal planetary environment. Once formed, the fireball products are widely dispersed over the planetary surface. This must have provided a major input of chemically reactive components on the Hadean surface, capable of providing the thermodynamic contrasts to support metabolism.

5.1 Impact-Produced Gases

Low-velocity impacts decompose volatile-bearing minerals to release H_2O, CO_2 and SO_2. Gerasimov et al. (1984) showed that impact vaporization thermally releases oxygen as O_2 and O into the plume. Though its lifetime would have been extremely short some oxygen may have escaped into the wider atmosphere. Carbon gases are also released in this process, mainly in the form of CO and CO_2 in a ratio of about $1:1$, close to the thermodynamic equilibrium ratio (Gerasimov et al. 2002). SO_2, COS, H_2S, and CS_2 also occur in impact processes (Ivanov et al. 1996; Gerasimov et al. 1994a, 1994b, 1997). Nitrogen is released from impacts mainly as N_2 and some oxides NO_x. Synthesis of HCN and traces of CH_3CN occurs. The amount of nitrogen in the impact-released gases is limited by its trace concentration in meteorites and surface rocks. More nitrogen can be involved in impact chemistry by interaction of a projectile or expanding plume with a nitrogen-containing atmosphere. Passage of a large asteroid through the Earth's atmosphere may in theory have produced nitrous oxides and hydrogen cyanide in an air shock wave. Acidic rains from dissolution of nitrous oxides in the water could be sufficient to increase ocean acidity and to perturb biological activity (Fegley et al. 1986; Prinn and Fegley 1987).

5.1.1 Redox State of Impact-Induced Gases

Gas mixtures formed after impacts are in disequilibrium compared to normal conditions. In addition to oxidized components the gases contain significant quantities of reduced components, including the simultaneous presence of H_2 and O_2, SO_2 and H_2S, CO_2 and CH_4. The most abundant reduced gases in quenched mixtures are H_2, H_2S, CH_4, and light hydrocarbons up to C_6H_6. Coexistence of O_2 and H_2 in high temperature silicate vapors is supported by thermodynamic calculations (Gerasimov 2002).

Iron is present as Fe^0, Fe^{+2}, and Fe^{+3} reflecting complex redox processes in the vapor. Formation of metallic iron is accompanied by the increase of the portion of Fe^{+3} compared to the starting sample. The main rock-forming element, silicon, has complex redox behavior, detected in Si^0, Si^{+2}, and Si^{+4} states (Yakovlev et al. 1993).

5.1.2 Formation of Organics During an Impact

Organic molecules can be synthesized in impact-induced vapors. Reported impact-generated hydrocarbons include: CH_4, C_2H_2, C_2H_4, C_2H_6, C_3H_4, C_3H_6, C_3H_8, C_4H_{2-8}, C_4H_{10}, and C_6H_6 (Mukhin et al. 1989). Production of unsaturated hydrocarbons is noticeably higher than of saturated. The only experimentally proven oxygen-containing organic molecule is likely to be CH_3CHO. The output of organics in and after an impact is correlated with the total amount of C and H in the starting material of the impact. The mechanism of hydrocarbon synthesis is still unclear. Possibly Fischer–Tropsch synthesis occurs (Zolotov and Shock 2000; Sekine et al. 2003), involving reaction of carbon monoxide and molecular hydrogen (abundant in the vapor). This reaction uses surfaces of condensing particles as catalysts and goes different ways depending on the type of catalysis (Chichibabin 1953)

$$n\mathrm{CO} + 2n\mathrm{H}_2 \xrightarrow{\mathrm{Ni,Co}} \mathrm{C}_n\mathrm{H}_{2n} + n\mathrm{H}_2\mathrm{O},$$

$$2n\mathrm{CO} + n\mathrm{H}_2 \xrightarrow{\mathrm{Fe}} \mathrm{C}_n\mathrm{H}_{2n} + n\mathrm{CO}_2.$$

The role of surface catalysis is also supported by observed synthesis of sufficient amount of complex kerogen-like organics bound to condensates, which were insoluble in solvents but detected by C–C and C–H bonding (Gerasimov et al. 2002).

5.1.3 Impact-Generated Silicate Vapor

Most of an impact-generated plume is silicate vapor. This mostly condenses into nanoscale particles during expansion and cooling of the plume (see Fig. 5). The cumulative mass of these particles is comparable to the mass of the projectile. Impacts on a scale similar to the putative K/T event would provide global dispersion of these particles, generating a dusty atmosphere, affecting global albedo and hence climate.

Analysis of experimentally produced condensates shows that the solids made in the experiment have structures that are highly out of equilibrium. Such disequilibrium products of late Hadean impacts may have provided a useful supply of reductants and oxidants to the first living communities.

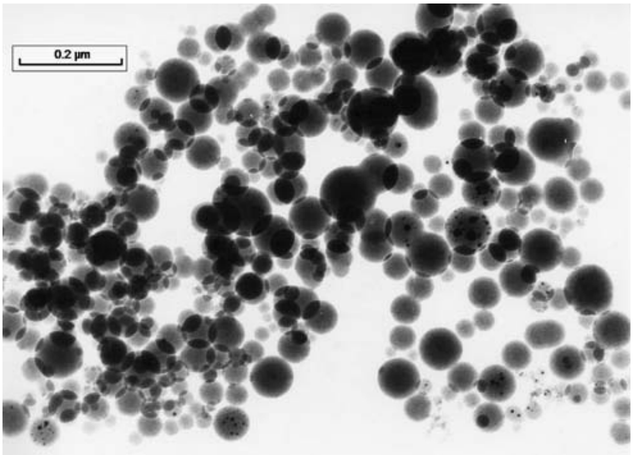


Fig. 5 Transmission electron microscope image of typical condensed particles. These are formed in an expanding vapor plume after high-temperature vaporization of a pyroxene (from Gerasimov et al. 1999)

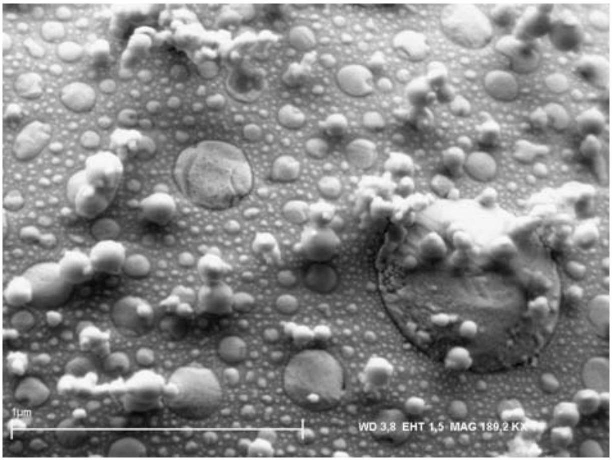


Fig. 6 Scanning electron microscope image of Fe-metal and FeS (*large*) liquates on the surface of ~9 µm glass spherule. From experiment on high-temperature pulse heating of carbonaceous chondrites (Murchison). The *scale bar* is 1 µm (from Yakovlev et al. 2002)

X-ray photo-electron spectroscopy analysis indicates the simultaneous presence of structures with isolated, chain, and framework polymerization of silicon–oxygen tetrahedrons (Gerasimov et al. 2002). Multivalence elements usually exist in collected condensates in all possible states. Redox behavior is also experimentally observed in impact condensates containing the main rock-forming element, silicon, which has been detected in Si^0, Si^{+2}, and Si^{+4} states.

Iron in impact condensates as Fe^0, Fe^{+2}, and Fe^{+3} may have been important in habitability by providing redox contrast. Metallic iron particles absorb siderophile elements and other reduced elements (Si, Ni, S, Mn, Pt, etc.). Such compositions in turn provide catalysts for various properties. Impact reduction of iron into a metallic state proceeds with formation of metallic nanoparticles. These are immiscible and segregate into a separate phase from the silicate melts . They can be dispersed into the vapor or found on the surface of silicate droplets (see Fig. 6).

6 Habitability Zones on the Surface: The Volume of the Oceans, the Emergence of Land

We now move from considering the inventory of materials to a discussion of the wider environmental constraints put on habitability by planetary tectonics.

Significant Freeboard, that is, the height which continents protrude from the ocean, is probably necessary for habitability. The oldest large body of rock, Greenland's Isua belt (up to $\sim$3.8 Ga), includes clastic sedimentary material—i.e. liquid (presumably water) was present. Unlike Venus, which has a unimodal relief, the Earth's surface is divided into continents and ocean basins. Sleep (2005) has pointed out the lucky coincidence that Earth, as always the Goldilocks planet, has just enough water that tectonics and erosion/deposition tend together to maintain continents that marginally rise above sea level. Continents are not inundated, nor are they immense and dry, as on Mars. Hynes (2001) showed a relationship between continental crustal thickness, continental crustal volume, and mantle temperature. Buoyancy of the oceanic lithosphere is controlled by the amount of melting at the ridges and mean age of the ocean floor (themselves functions of mantle temperature).

Galer and Mezger (1998) investigated the antagonistic relationship between uplift and rifting processes driven by tectonics and the erosion and deposition of the sedimentary cycle. They inferred that the balance between the two was the main mechanism that kept continental freeboard constant, by regulating continental thickness. From an assessment of the geological settings of ten relatively undisturbed greenstone belts, they inferred that at 3 Ga ago the mean continental thickness was $\sim$46 km at the time of crustal stabilization, and the mean thickness of oceanic crust was around 14 ± 2 km. This thickness of oceanic crust tallies with evidence for a somewhat hotter mid-Archean potential temperature of the mantle (Nisbet et al. 1993).

As far as habitability is concerned, it is clear that some continents have protruded from the Earth's oceans ever since the mid-Hadean (Nisbet 1987). In searching other planets, this may be a useful though not perhaps necessary criterion: that there is freeboard.

7 The Small Scale—From the Volcano to the Ion: Supply of Key Components for Biochemistry

Life is small, especially single-celled microbial life. Habitability is defined on the scale of a cell, not a planet. A planet is only habitable if it is habitable at the very smallest scale. There must be a supply of key components that is accessible to life that is restricted in movement on the micron scale.

At the chemical core of biochemistry are the metal centers of the metalloenzymes. These are 'house-keeping' proteins, many of which may have origins in the very greatest antiquity. Iron in particular is central to a wide array of enzymes that appear to be of very great antiquity. Typically Fe is in association with S. The iron enzymes include catalases, peroxidases, ferredoxins, oxidases, and all nitrogenases. Nickel examples include enzymes such as the hydrogenases (used in dealing with hydrogen and probably very ancient) and urease (essential to the nitrogen cycle). Carbon monoxide dehydrogenase, which is at the center of the acetyl-coA metabolic pathway, contains nickel, zinc, iron and molybdenum. Methanogens use coenzyme F_{430} and hydrogenase, both nickel enzymes: thus nickel is essential to methanogenesis. Nitrogenase (at the core of the nitrogen cycle), is an Fe-Mo enzyme. A particularly interesting class of enzymes includes metal–4N groups. The best known of these include chlorophyll and bacteriochlorophyll, which have Mg at the center, and haem, with Fe. Metal-4N groups occur in alkaline conditions, such as those in hydrothermal systems around komatiites.

The presence of Mo in sediment has been seen as evidence for oxic waters in deep time (Siebert et al. 2005), as oxic conditions are needed to mobilize the Mo. There that there are other ways to mobilize Mo: thiosulfate, which is plausible in Archean oceans, will mobilize reduced Mo. Arsenic resistance may be ancient (Gihring et al. 2003) and may even pre-date the last common ancestor.

All these house-keeping proteins suggest a geological origin where the metals and S would be in abundant supply. The first life could not seek out metals in the way modern life does. The metals must have intruded into the organism: i.e. for the 'unskilled' organism to incorporate metal ions, the setting must have had abundant and accessible supplies, such that entry of metal into the cell was inevitable, not a rare accident. The most probable setting is in and around hydrothermal systems. Fe–S, for example, immediately suggests hydrothermal settings, as does the Fe–Mo association. Moreover, the commonness of metal–S groups immediately suggests sulphides to a geologist: hydrothermal settings.

The high potential temperature of the Hadean mantle is relevant here. If the temperature of the mantle in the late Hadean was, say, around 1800 K, then komatiitic volcanism would be extremely common. Komatiites are highly magnesian: they are very hot; they contain abundant Ni. During submarine eruptions, hydrothermal activity probably begins very early, as a komatiite lava flows, creating Ni–S rich layers in the cooling lava. The high MgO content means that the fluids can be very alkaline. We speculate that Ni-enzymes and also Mg–4N complexes first evolved in such settings. Had the mantle been cooler (i.e. no komatiites), then such settings would not have been possible. In other words, the high potential temperature and the consequent presence of komatiites may have made the Earth much more habitable than a similar but colder planet, more conducive to the origin of life as we know than the cooler Earth we live on now.

Saito et al. (2003) have studied the trace metal preferences of cyanobacteria. They showed that trace metal preferences in cyanobacteria are consistent with their evolution in a sulphidic ocean setting. Saito et al. (2003) suggested a variant on the hydrothermal hypothesis put forward above. They pointed out that in an early anoxic ocean, dominated by high concentrations of Fe, and with abundant Fe supply, and dissolved sulphidic species, the relative availability of trace metals would have been similar to the availability in a sulphidic system. In order, the availability would have been Fe $>$ Mn, Ni, Co $\gg$ Cd, Zn, Cu. Saito et al. (2003) suggested it is possible that strong aqueous metal-sulphide complexes were as important as mineral precipitation in incorporating metal enzymes into biology. Possibly marine geochemistry and marine biology have co-evolved.

8 The Small Muddy Doorway of Habitability

There must have been a pre-biotic 'doorway' (not a window, but an entry point) of habitability, when life could begin. But this doorway would have been small. It would only have been open where there was local thermodynamic disequilibrium, created by geological and atmospheric/space contrasts.

Figures 7 and 8 illustrate this. The late Hadean surface environment would typically support only small and local redox contrasts. Volcanisms would have been prolific, given the mantle's high potential temperature. Volatiles such as SO_x that were degassed from volcanoes would have been oxidized in an atmosphere that was photolytically losing H to space: this would provide a somewhat more oxidized ocean. Lavas erupting would provide a

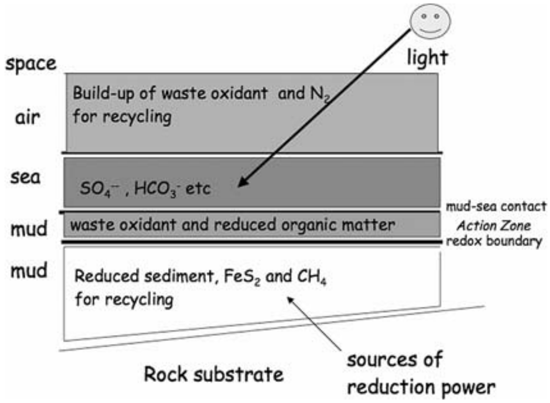


Fig. 7 Mud—the habitat

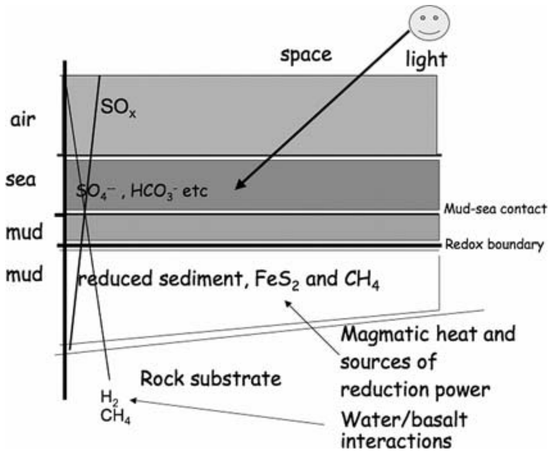


Fig. 8 Thermodynamic contrast before life. The atmosphere would have been marginally more oxidized (because of H escape to space and more generally from photolysis) while the interior would be more reduced

somewhat more reduced geological substrate. Methane haze and CO from photolysis would also be accessible.

A major early source of redox contrast may have been supplied by the hot lava via the intermediary of H_2. When lava and seawater interact, hydrolysis of ferrous oxide components in rocks produces hydrogen, e.g.:

$$3Fe_2SiO_4 + 2H_2O = 2Fe_3O_4 + 3SiO_2 + 2H_2.$$

The liberation of H_2 is always associated with formation of magnetite as this phase stabilizes ferric iron at low redox potentials. Epidote too does this. Hydrothermal systems similarly can produce inorganic methane.

The crucial site of contrast would be the ocean/mud interface, especially in and around hydrothermal systems. 'Mud' is the clay of life. In the Hadean, muds would have been sparse and magnesian–saponitic and nontronitic, not aluminous kaolins. Perhaps it was in the alkaline muds around hydrothermal vents from a recent seafloor komatiite flow, that key early housekeeping metal–4N proteins like the cytochrome family first formed. In such settings, brucite would scavenge borate and phosphate from seawater. Abiotic formation of pentoses, particularly ribose, would be promoted by the high pH (Holm et al. 2004) and boron (Ricardo et al. 2004).

The presence of redox contrast between different parts of the habitat is essential. An ongoing contrast is only sustainable if geological processes continually introduce supplies of relatively oxidized and relatively reduced species. The supply of reduced species in a hydrothermal system, especially one that involved very hot ultramafic rocks, would be abundant. Both H_2 and CH_4 would be likely. However, the supply of relatively oxidized material must depend on photons. Oxidized species would originate in the air. They would have to exchange with ocean waters by rainfall and wave bubbles, and then the supply to the putative early cells would depend on the interaction between water currents and the water/mud interface.

Sulfate, thiosulfate, elemental sulfur, and perhaps carbonyl sulfide (OCS) would be the most likely carriers of redox contrasts. They would all rain out of the atmosphere into the ocean. Both thiosulfate and OCS can help to mobilize metals. On the modern Earth, sulfur in arc volcanoes is in part from subduction of sulfide originally precipitated in the ocean crust by hydrothermal circulation, which in turn derived the S from sulfate. In the modern ocean the sulfate is mostly in this oxidation state because of photosynthesis. In the pre-photosynthetic planet, sulfate would thus not necessarily be so readily available.

9 Exploiting the Muds: Widening the Habitability of the Earliest Archean Ecosphere

Life makes its own habitat. One of the many paradoxes of life is that the planet is habitable because it is inhabited. Today the atmosphere is an almost wholly biological construct: the nitrogen fluxed by nitrifying and denitrifying bacteria, the oxygen and carbon dioxide cycled by photosynthesis and respiration, the methane by methanogens and methanotrophs. Even the abiotic argon is arguably biologically controlled, as it is brought to the surface via granitoid magmas, that need subducted water, and water is stabilised on the planetary surface by the biological thermostat.

The arrival of life immediately makes the planet change: life generates its own habitability. This is because life in one niche of habitability immediately creates new niches of habitability. Imagine, for example, a tiny population of first living cells existing in a small vesicle in porous rock on the outer edges of a hydrothermal system, exploiting a supply of H_2 from water reacting with the rock, or using the redox contrast between a) incoming fluxes of slightly more reduced S species in the hydrothermal fluids and b) intermittent inputs of slightly more oxidized ambient seawaters. For the first few generations, the population will expand until the local habitat's resources are exploited. But there will be much cell death, either in 'drought' periods when one or the other inputs is absent for too long, or simply because of the Malthusian corrective that the population expands until some cells are driven to the edge.

Once there is cell death, there is reduced carbon lying on the floor—a new niche of habitability. There will be hydrogen coming up from the water-rock system below: thus there will be a new niche of habitability, using H_2 to turn the dead bodies into methane. In the air, methane haze and CO from photolysis provide substrate for life. The result of life's activities is to focus the redox boundaries more sharply: that in turn defines the zone of habitability and increases the productivity of the system, which goes from a disseminated one-niche to a set-apart two-niche system. Dead bodies foster life.

Sustained habitability makes sophisticated use of dead bodies. Like the ancient Greek hero Trygaeus, who saved Peace by riding to the Moon mounted backwards on a dung beetle, which he fuelled himself, life rises to the heavens on its own excreta. The use of hydrogen to recycle dead bodies made of organic carbon compounds will generate methane. Methane

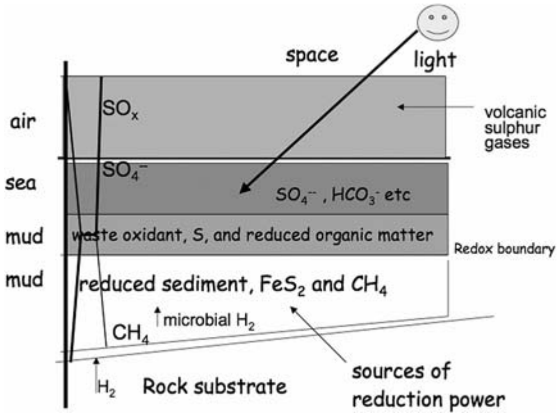


Fig. 9 Remaking the habitability of the mud, before photosynthesis. Even before photosynthesis began, life would begin to rearrange redox gradients in the mud. Dead organic matter would collect in the lower mud, making it more reducing (i.e. more CH_4 and H_2S, less SO_4) while upward release of more oxidized species such as CO_2 would make the upper environment more oxic. Microbial organisms typically focus redox contrast, and create a sharply defined zone where life flourishes

in turn is a fuel to consortia of cells that can use it to reduce sulfate: a new niche arises. By early Archean time, perhaps in the interval 4 Ga to 3.8 Ga ago, before the Isua belt was laid down, the pre-photosynthetic community could likely have created a very diverse set of niches, remaking the planet to be far more habitable than before life (Fig. 9). This biosphere would have been non-photosynthetic, depending purely on available redox contrast. In other words, the habitable part of the planet would have been limited to the close proximity of volcanic hydrothermal systems.

10 Habitability of Highly Mineralizing Environments

There is a potential paradox here: on one hand, for all the reasons above, volcanic hydrothermal systems were probably the first habitable zones on the early Earth; on the other hand, they are dangerous for life, because of the lethal effects of mineral precipitation on cells. Fossilization of microbial cells possibly occurs by precipitation of many different minerals: e.g. silica (e.g. Westall et al. 1995; Toporski et al. 2002; Benning et al. 2004), manganese oxide (Tebo et al. 2004), calcium phosphate (Benzerara et al. 2004a, 2004b). Cells can be totally entombed in the precipitates (Benning et al. 2004; Benzerara et al. 2004a, 2004b).

The ecological and evolutionary implications of fossilization are profound (see Caldwell and Caldwell 2004 for a conceptual view of these issues). Mineralization processes have a major influence on habitability. The traditional view considers microbes to be purely "*passive*" in the precipitation of the minerals: microbes modify indirectly the chemical conditions of the surrounding environment by their metabolic activity and hence foster mineral precipitation. This is substantiated by chemical/thermodynamical modeling (e.g. Frankel and Bazylinski 2003).

It is possible that the ability to biomineralize may only be a side-effect of metabolism. It does not provide a selective advantage to a microbe. It may actually be disadvantageous: the precipitation of few hundred nanometer thick layers of minerals around cells may limit diffusion of nutrients necessary for life; the formation of nanometer-sized crystals may disrupt cellular structures and hence be lethal. The fitnesses (i.e. an individual's ability to propagate its genes) of a microbe promoting biomineralization vs. that of a microbe inhibiting biomineralization have, however, never been measured.

Some strategies developed by microbes to inhibit precipitation of minerals on their membrane might be operating in highly mineralized environments. One example is the formation of sheaths, which are extracellular tubes surrounding microbes and that offer preferential nucleation sites for crystals. Microbial cells can get rid of these tubes and form new sheaths (Phoenix et al. 2000; Konhauser et al. 2001). Emerson and Ghiorse (1992) showed that sheathless variants arise spontaneously in laboratory cultures if predation and mineral precipitation are no longer present. Schultze-Lam et al. (1992) proposed that some cyanobacteria synthesize protein surface layers (S-layers), which provide nucleation sites for calcite precipitation and that they can shed when encrusted in mineral precipitates.

There is however no conclusive evidence that those strategies are more developed in microbes inhabiting highly mineralizing environments than anywhere else. Moreover, the comparison of the various studies on mineral precipitation on bacterial cells is confusing regarding whether this process is lethal or not (see Kappler et al. 2005). Some authors suggest that biomineralization on cell walls occurs during or after death of cells (e.g. Wierzchos et al. 2005), while others report the existence of viable cells encrusted by minerals (e.g. Phoenix et al. 2001; Tebo et al. 2004).

Seemingly passive biomineralization may actually provide an advantage to microorganisms. First of all, some studies contest that mineral precipitation on cells is always a passive mechanism (Castanier et al. 1999). They suggest that some species have a better ability than others to precipitate calcium phosphates or calcium carbonates (Castanier et al. 1999). Could this ability to precipitate minerals be advantageous in some cases?

Several potential advantages have been proposed for extracellular biomineralization: detoxification of toxic heavy metals, reactive oxygen species, UV light, predation or viruses; protection against immune system for pathogenic microbes; protection against grazing; storage of an electron acceptor for later use in anaerobic respiration; scavenging of micronutrient trace metals (e.g. Sommer et al. 2003; Mire et al. 2004; Tebo et al. 2004; Ghiorse 1984). Chan et al. (2004) suggested that extracellular iron oxide precipitation may provide energy to microbial cells. Whereas this has almost never been proposed for minerals like silica, calcium carbonates or calcium phosphates, it is usually considered that lead phosphate precipitation by bacteria is a detoxification process. Microbially driven calcium phosphate precipitation, though, shares many similarities with biomineralization of lead phosphate. It is thus reasonable to consider that microbial calcium phosphate precipitation may have a similar ecological 'status' as lead phosphate precipitation and such biomineralization processes may have played a role in the habitability of highly mineralizing seemingly "toxic" environments.

11 Sub-Surface Habitability

Modern marine microbial life does not only live in the water and water/mud interface. It is prolific down to several kilometers below the terrestrial surface and the seafloor. This is a large habitat, that preserved a wide anoxic environment even when the air became oxic.

The limit for life essentially is given by the increasing temperature and probably corresponds broadly to the 120°C isotherm (Kashevi and Lovley 2003) or, in the case of Earth, a depth of 2–12 km (typically 4 km), depending on the geotectonic setting. Both sediments (Aitken et al. 2004; D'Hondt et al. 2002, 2004; Parkes et al. 1994, 2005; Pedersen 2000) and igneous rocks (Moser et al. 2005; Pedersen 2000; Stevens 1997; Stevens and Mckinley 1995; Stevens et al. 1993) are inhabited by microbes. Colonized voids may range from microscopic pores to caves (Boston et al. 2001) and include active faults (Moser et al. 2005).

In muds and sediments undergoing diagenesis, very abundant microbial life occurs, living by exploiting the buried archive of more and less-reduced material: reacting dead organic matter with sulfate, for example, in processes such as anaerobic methane oxidation.

In igneous rocks, microbial colonization has not just been documented as recent activity but there is also evidence of a fossil record of subsurface life (Furnes et al. 1999; Furnes and Staudigel 1999; Hofmann and Farmer 2000; Kretzschmar 1982; Mckinley et al. 2000; Trewin and Knoll 1999). There is free energy in basaltic glass relative to crystals. It has been proposed that endolithic microorganisms can form long channels in basaltic glass and use the oxidation of Fe(II) as an energy source (e.g. Furnes et al. 2005). Although the biogenicity of those structures is debated (Brasier et al. 2005), it is ascertained that microbes can promote dissolution and hence affect habitability (Rogers and Bennett 2004; Benzerara et al. 2005).

Compared to surface environments on Earth, the subsurface microbial biomass has very low productivity, but can be very long-lived and cell populations can be abundant (though many orders of magnitude less than on the surface). Most subsurface life is indirectly dependent on the surrounding inventory of surface derived oxidants or reduced organic debris. Even in the case of absence of surface-derived oxidants, microbial life based on methanogenesis using abiotically produced hydrogen (from water-rock interactions) and CO_2 is possible (Stevens and Mckinley 1995; Chapelle et al. 2002). It has been estimated (Pedersen 2000) that the subsurface biomass is similar to the combined continental and marine biomass. On a planet subject to surface catastrophe—for example a cometary or meteorite impact, or undergoing a snowball, the subsurface community provides habitability that protects the long-term continuity of life. Also, where surface conditions have degraded over time (e.g. Mars), the subsurface may still provide a water-rich (Burr et al. 2002; Clifford and Parker 2001) habitat for microbes long after the demise of all surface life. Moreover, peroxide reacting with iron or relict organic matter can provide a photolysis-based niche (e.g. on Mars).

The crucial factor for subsurface life is the presence of suitable redox couples to maintain a source of energy. Since the very action of microbial life will, in a closed system, eventually lead to a state close to chemical equilibrium, an input of oxidants from outside is required (Weiss et al. 2000). In sediments and fractured igneous rocks, diffusion and/or advective flow of water containing surface derived oxidants is required. This, in addition to temperature, is the limiting factor for subsurface life. However, non-terrestrial sources of redox contrast will in general be different from those on Earth. Here, even the mid-ocean ridge black smoker communities are dependent on surface oxygenic photosynthesis (which provides free oxygen and sulfate to seawater, the main oxidants reaching the hydrothermal community).

On a planet without oxygenic photosynthesis, or even without an ocean, redox contrast would be harder to find but may be related to photolysis reactions in the atmosphere as in the case of early Earth. The Martian subsurface is an obvious candidate to host such settings with accessible redox contrast, for example where large units of methane clathrate (produced abiotically by water-rock reactions with hot lava) are in proximity with oxic waters circulating to the surface.

Subsurface life on Earth is very ancient. Buried carbon and sulfur in the Belingwe belt (Grassineau et al. 2006) has probably gone through microbial processing during diagenesis. It is likely that subsurface sediments and igneous rocks have hosted microbial life, including anaerobic methane oxidizers, ever since the early-mid-Archean (Westall et al. 2006aa). On present Earth, microbial life is ubiquitous in sediments to depths of hundreds of meters and in igneous rocks to several km.

Subsurface habitats likely exist on other planets. Even where the surface is barely habitable, subsurface environments may be suitable for microbial life in a similar manner as on

Earth. The deep terrestrial subsurface is relatively good analogue for such subsurface habitats on other planetary bodies, such as Mars. Deep fluids in both cases are low in oxygen and largely controlled by water–rock interaction. The major difference is the presence of abundant oxidants at the Earth's surface, while the deep Martian subsurface is largely sealed from surface water circulation by permafrost. Besides the possibility of present subsurface life on other planetary bodies, the palaeosubsurface is a promising area to search for extinct life. Any former cavity in a rock must be considered a potential palaeohabitat. Non-sedimentary rock formations, especially where signs of aqueous alteration are present, must not be neglected in the search for former life. Suitable biosignatures allowing the recognition of such ancient subsurface life may be of morphological (well-preserved filaments) or geochemical nature (e.g. evidence of low-T sulfate reduction).

In this context, the discovery of methane in the Martian atmosphere (Formisano et al. 2004) is interesting. It could be a microbial waste product. However, an abiogenic source is also plausible—formed, possibly many millions of years ago, by hydrothermal systems around volcanism, and recently sweated out of clathrates by seasonal cycling. Given the ubiquitous oxidation of the Martian surface, sites of methane venting, whether the gas is of abiotic or biogenic origin, could support a marginal community of methane oxidizers.

12 The Intangible Air, Connector of Habitability: The Methane Greenhouse

In particular, methanogens would have been important in the pre-photosynthetic Earth. The methane would have had no obvious early biological sink (except perhaps anaerobic methane oxidation, but limited by sulfate supply), and methane would thus have collected in the atmosphere. Methane would have warmed the planet. Just possibly, the Earth is habitable because it is inhabited: possibly the methane produced by early life saved us from an eternal snowball (Pavlov et al. 2001).

Such a world would sequester large amounts of methane in clathrate hydrate in icy sediments. Large sequestrations can cause burst-out catastrophes: e.g. where a volcano erupted over a major clathrate-trapped pool of free gas. These very large methane bursts could then trigger global warming and release of the other clathrates (as may have taken place more recently, in the Paleocene/Eocene thermal maximum).

Over the eons, the carbon cycle has continuously released oxidizing power (Hayes and Waldbauer 2006). Most of this oxidizing power is now represented by Fe^{3+} that has accumulated in the crust or been returned to the mantle via subduction, but about 3% is the oxidized atmosphere (Hayes and Waldbauer 2006) and the rest sulfate. Nevertheless, in the sediments methanogens can still dominate.

Walker (1987) suggested that at times the Archean biosphere was 'inverted', with the atmosphere more reduced than the muds (Fig. 10). In the long run, the contrast between the more-reduced mantle and the more oxidized air, subject to solar photons and hydrogen escape to space (Catling et al. 2001), would restore the normal state-of-affairs, but potentially, Walker-world inversion events could have been sustained for long periods, when for some time the methane-rich atmosphere was more reduced than the mud. Such events would trigger major evolutionary changes. Mass extinction is unlikely in a microbial planet where, via winds and ocean currents, even one surviving cell can repopulate a niche very rapidly, but major changes such as a redox inversion could spur very rapid evolution into new niches.

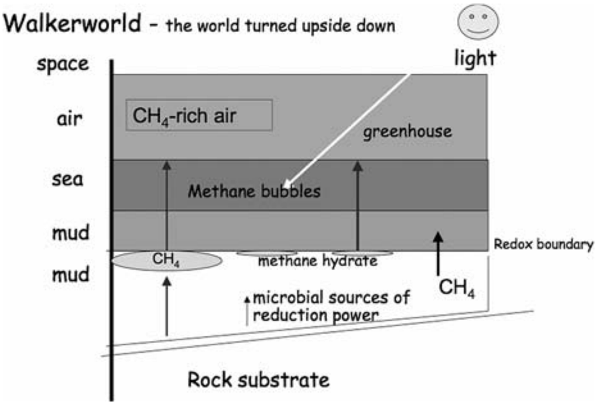


Fig. 10 Walkerworld—the upside-down biosphere. From time to time in the Archean, catastrophic release of methane from mud may have overwhelmed the slight oxidation reservoir of the atmosphere, creating an environment in which the atmosphere was more reducing than the magmagenic interior of the planet

13 Reworking the Habitable Mud: Anoxygenic and Oxygenic Photosynthesis, and Burrowing Metazoa

The history of the changes imposed by photosynthesis is beyond the scope of this chapter. Nevertheless, a brief comment is tempting. The evolution of *anoxygenic* photosynthesis, perhaps prior to 3.8 Ga ago, processing species such as H_2, H_2S, NH_3, CH_4, FeO, etc., would have vastly increased the habitability options. By 3.4 Ga ago, photosynthesis was well-established (Tice and Lowe 2004; Westall et al. 2006aa) presumably anoxygenic as the atmosphere was anoxic. Compared to today, the air was relatively CO_2-rich 3.2 billion years ago (Hessler et al. 2004).

Oxygenic photosynthesis, when it arrived, was a much more dangerous proposition than anoxygenic light-capturing. Waste oxygen made life banished anaerobes from the surface, and risked freezing the planet by destroying the methane greenhouse. This would have reduced habitability briefly. In the longer run, the vast increase in productivity (Kharecha et al. 2005) would create a much greater habitat, both on the seafloor and in the water.

The arrival of the nematode would have been the next dramatic change in habitability. By burrowing and fluxing fluid through the mud, the presence of worms very greatly expands the productivity and hence habitability of the sea floor (Fig. 11). Though the overall range of habitability would probably not have changed much (bacteria are deep in the sediment), the fluxing of nutrient and redox power between sea and mud would have increased markedly. Although microbial activity extends deep into the sediment column, on a microbial world photosynthesis and hence the associated major zone of high productivity is confined to a thin skin a few millimeters thick on an otherwise dead planet. In contrast, a biosphere with worms enlarges the high-productivity surface-linked community by pumping fluids around. The zone of prosperity becomes several meters thick, or more in good muds: the production of the zone of habitability inflates like the universe itself. Of course, the burial biosphere is much deeper, as much as several kilometers thick (and tens or hundreds of kilometers wide) in sedimentary basins filled with cool organic-rich, but here productivity is much lower, even if cell counts are high.

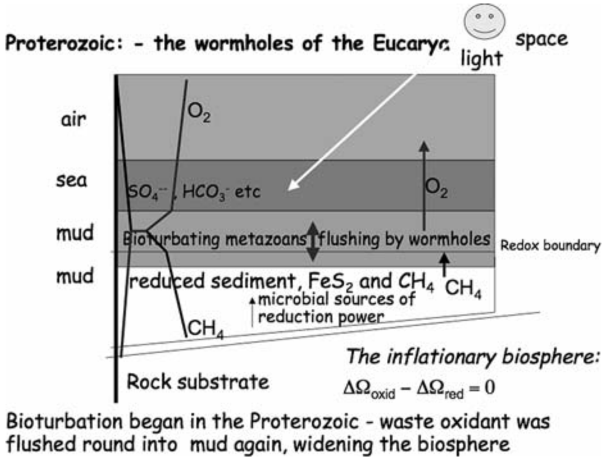


Fig. 11 Wormholes—the inflation of habitability. Where worms pump fluid up and down in the sediment, the access of deeper microbial consortia to oxidant is much improved. Cycling of chemical species is enhanced. This process creates a more productive deeper (reduced) zone and a matchingly larger more oxic zone. The interface is more focussed and there the redox gradient becomes sharper. The net redox sum remains the same, but the two opposite reservoirs increase

14 Habitability of Another World—Is There a Rock Cycle on Mars that Can Sustain Life?

Minerals and rocks are stable only under the conditions at which they form, though kinetics and activation energies permit sustained disequilibrium (if not, diamond rings would vaporise to CO_2). Changing these conditions will initiate metamorphism of the rock and its minerals. Therein is disequilibrium, and resources for life. On Earth plate tectonics regularly remodels the face of the planet and is central to the sustained supply of ions and disequilibrium for biogeochemistry. Is there also evidence for plate tectonics and related disequilibrium among surface rocks on Mars?

The discovery of magnetic lineation in the southern Martian highlands (Connerney et al. 1999) suggested that Mars experienced an early era of plate tectonics, forming surface material by crustal spreading in the presence of a reversing dynamo (Connerney et al. 2005). Sleep (1994) argues that tectonic features on Mars such as Gordii Dorsum resemble plate break-up margins and that the Tharsis Montes volcanoes, which are located at the south to north dichotomy boundary began as volcanic centers along a Martian subduction zone. This suggests plate tectonics produced the northern lowlands of Mars. Possibly on early Mars plates of cooled basaltic material might have subducted into a mafic-gabbro mush mantle. Water release would then cause voluminous tonalitic (TTG) melts.

The magnetic lineation is correlated with the oldest surface units, implying the dynamo's demise very early in the Martian evolution (Nimmo and Stevenson 2000). Yet there is evidence (Head et al. 2001; Frey et al. 2002) suggesting considerable crust formation after the inferred end of plate tectonics at about 4 Ga ago. However, early-ceszation plate tectonics models are difficult to reconcile with such substantial formation of post plate tectonics crust. The more efficient the cooling, and the more thermal power is available for the dynamo, the less likely is the production of crust after ceszation of the plate tectonics regime (Breuer and Spohn 2003).

Van Thienen et al. (2004), using buoyancy arguments, suggest plate tectonics was unlikely during early Martian history. There is only a relatively low operational temperature window for plates (1300–1400°C), outside which plates do not attain enough negative buoyancy to subduct on Mars, within reasonable time scales.

The lower gravity of Mars allows deeper hydrothermal circulation through cracks and hence more hydration of oceanic crust so that water is more easily subducted than on the

Earth (Sleep 1994). However, the chemistry of the SNC meteorites has been interpreted as implying Martian mantle has been dry since core formation (Dreibus and Wänke 1987, 1989; Carr and Wänke 1992). On the other hand, cooled basaltic material might subduct into a mafic-gabbro magma ocean and due to lower gradient with depth on Mars slabs could go down hundreds of kilometers (Warren 1993; Bridges and Warren 2006). The subducted water might be responsible for extensive volcanic eruptions.

Lenardice et al. (2004) postulated a model tectonic history of Mars. In this, Mars initially underwent active lid tectonics, driving a geodynamo and producing the southern highlands through some mixture of melt generation and crustal accretion. The growth of the crust increased the mantle temperature, lead to the ceszation of plate tectonics, with the tectonic style transforming into a stagnant lid regime. If the model is correct, plate tectonics was active, if at all, only for a very short period of less than 300 Million years at the beginning of the planet's history.

Evidence for rock diversity on Mars resulting from recycling induced by plate tectonics is hard to find—yet this is central to habitability. If Mars did not cycle its chemistry, it would soon have become uninhabitable, an unsuitable substrate, when life was starved of nutrient supply.

14.1 Martian Volcanism

The bulk composition of the SNCs as well as the bulk composition of the Martian soil is basaltic (McSween 1994; Rieder et al. 1997a, 2004; McSween et al. 2004; Mustard et al. 2005; Yen et al. 2005) with no evidence for alterations induced by subduction. Some rocks at the Mars Pathfinder landing site are characterized by higher SiO_2 contents $>55\%$ (Rieder et al. 1997b; McSween et al. 1999) indicating a more andesitic composition (although the measurement of higher SiO_2 composition could also be due to some form of alteration rind on the rocks, similar to desert varnish). On Earth, icelandites originate from fractionated crystallization of basaltic magma in rifting anorogenic environment. The composition of Martian andesitic basalts is broadly comparable to icelandites (McSween et al. 1999). Thus, the composition of Martian basalts shows no direct evidence of formation under aqueous conditions, but the high SiO_2 content may indicate a volatile rich basaltic component. However, Bridges and Warren (2006) report on the evidence of water in the evolution of shergottites give by their lithium chemistry and Dann et al. (2001) showed, that crystallization of shergottitic melts under water saturated conditions can lead to Fe and Al_2O_3 depletion and SiO_2 enrichment to produce andesitic-like melts. Re-melting of hydrated mafic rock in a mush magma ocean might have supported these processes.

Thus, although there is no clear evidence for subduction, there is abundant evidence of volcanism that could have supported hydrothermal systems that were habitable.

15 Water-Rich Environments on Mars

Volatiles are crucial to habitability (see Chap. 4 of this book). Not only are they essential for life, they also shape the enviroment. Although we do not have evidence for a rock cycle driven by plate tectonics, alteration of rocks due to weathering and transportation is common on Mars. The surface geology exhibits giant outflow channels which start fully developed at discrete sources and expand for hundreds of kilometers. (e.g. Sharp and Malin 1975; Baker et al. 1992; Carr 1996; Jaumann et al. 2002) branching down-stream and having various streamlined bedforms on their floors. Valley networks are smaller but nevertheless show fluvial characteristics. Their origin is drainage dependent (e.g. Sharp and Malin 1975; Pieri 1980; Carr 1996; Jaumann et al. 2002), with various glacial and periglacial features (e.g. Squyres and Carr 1986; Kargel and Storm 1992; Neukum et al. 2004). There is evidence for continuing activity (Malin et al. 2006).

Erosion on Mars was most likely driven by wind and water, either in liquid form or as ice. Fluviatile phases are old and date back mostly to the Noachian and early Hesperian times (e.g. Baker et al. 1992; Carr 1996; Jaumann et al. 2005). Glacial activities are ongoing, even in younger Amazonian times. In parts they may be recent (Neukum et al. 2004; Head et al. 2005; Hauber et al. 2005b). Polar layered deposits occur. Mars had and has a hydrologic cycle that varied in intensity with time. However, compared to Earth the erosion on Mars is not well developed as indicated by the relatively rare distribution of fluvial and glacial features and low erosion rates (e.g. Arvidson et al. 1979; Carddock and Maxwell 1993; Golombek and Bridges 2000; Jaumann et al. 2005), which are a few orders of magnitude lower than on Earth. Nevertheless fluvial and glacial erosion induces a rock cycle driven by sediment production through weathering followed by transportation and re-deposition (Fig. 12).

Aeolian processes also contribute to erosion, by cycling the finest fraction of material. Mechanical weathering produces dust and sand-like material, which is redistributed by wind and deposited in dunes (Fig. 13).

Chemical alteration of rocks is found only in localized concentrations of hydrated phyllosilicates and sulfates such as gypsum and kieserite (Bibring et al. 2005; Langevin et al. 2005; Arvidson et al. 2005; Gendrin et al. 2005), mostly in northern circumpolar regions, layered terrain in Valles Marineris and in Sinus Meridiani, indicating aqueous environments but a minor role of water in modifying surface rocks. Due to the high cation to sulfur ratio in typical mafic rocks the sulfur needs to oxidize rapidly compared to clay formation so that CaO and NaO do not neutralize the acid. Therefore these environments are weak biomarkers, as is jarosite.

Sedimentary rocks are exposed in the Meridiani region at the Mars Exploration Rover Opportunity landing site. These sediments are flat lying, finely laminated with fine-scale cross lamination in some areas. They are sulfur rich and contain abundant sulfate salts (Squyres et al. 2004). These rocks are thought to be a mixture of chemical and siliciclastic sediments formed by episodic inundation of the shallow surface by water in combination with evaporation and desiccation probably under acid-sulfate conditions (Klingelhöfer et al. 2004; Squyres et al. 2004). Embedded hematite-rich concretions within these sediments also indicate the involvement of water. Sulfur, chlorine and bromine indicate chemical mobility and separation during a period of aqueous activity (Haskin et al. 2005; Gellert et al. 2004). Although there is strong evidence for chemical alteration of rocks at the Meridiani site, other places on Mars, particularly the Pathfinder and Gusev sites, indicate that physical weathering might have played a more significant role than chemical weathering in the rock alteration process on Mars (Morris et al. 2005).

The thin Martian atmosphere is not able to hold much water (less than 10 μm) vapor and thus is a very water-poor environment. Current Martian climatic conditions of about 6 mbar and seasonal average temperatures below freezing, prohibit water on the surface in most Martian surface regions and for the most time of the year. Small amounts of water may be feasible on the surface during summer daytime temperatures and in low laying areas, especially when perihelion occurs in Mars's northern hemisphere. That said, today's water occurs mostly in the atmosphere and at the poles. The water content in the atmosphere is about 1–2 km^3, which is equivalent to a 10 μm global water layer. One meter of equivalent global water layer translates to 1.44×10^5 km^3. In contrast, the water in the Earth's atmosphere would cover Mars with a global layer of 10 cm.

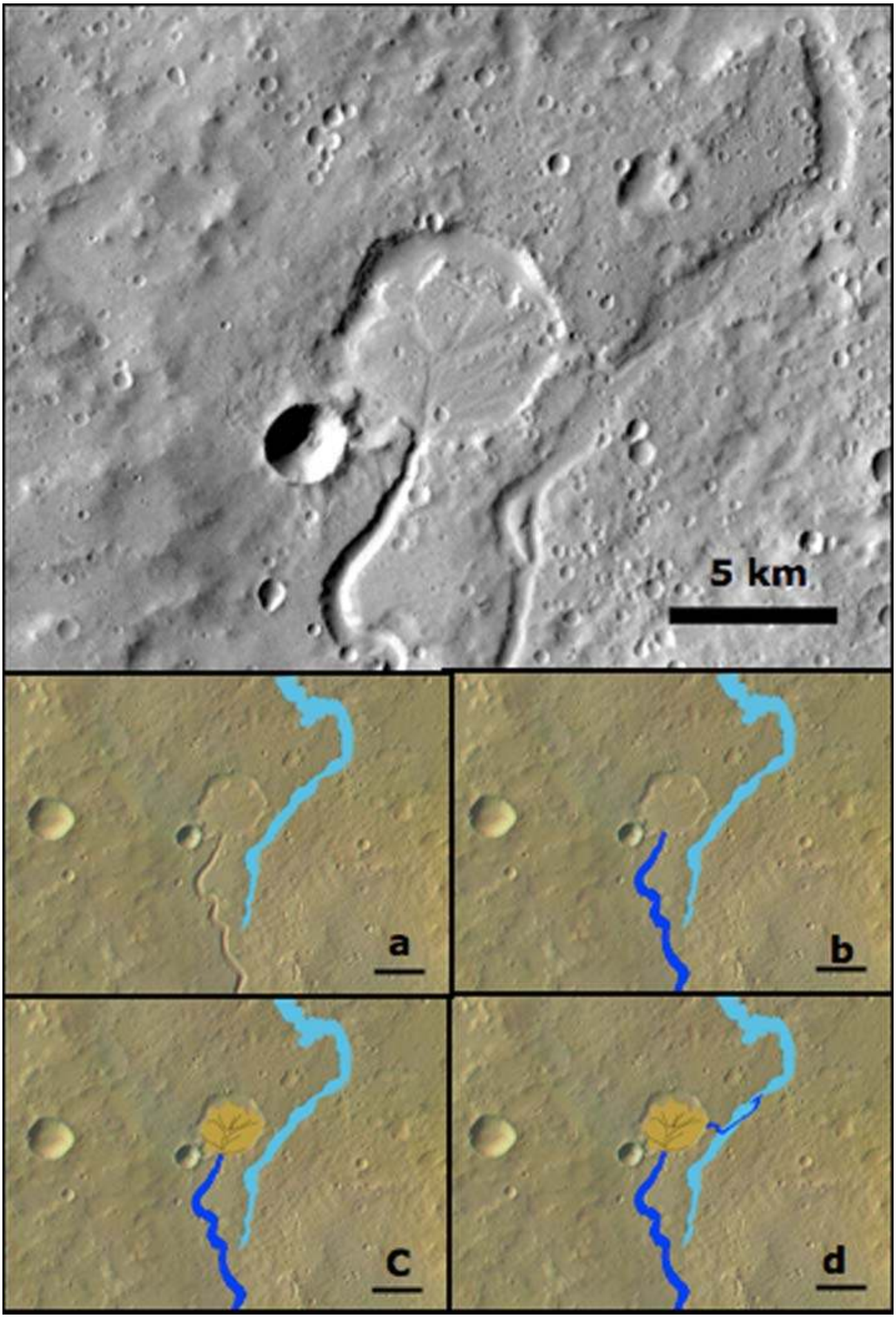
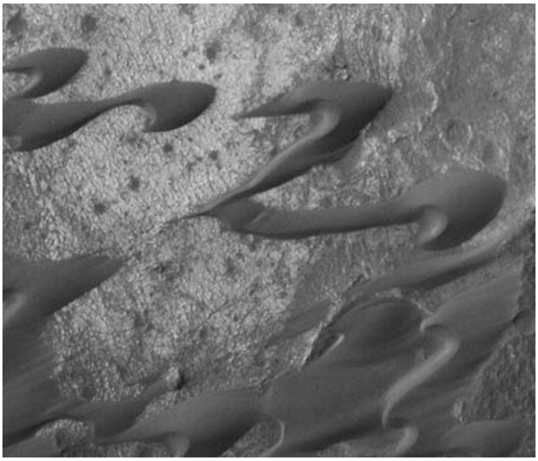


Fig. 12 Valley and delta in Xanthe Terra on Mars (Hauber et al. 2005a). Remnants of an eroded older channel (**a**) are superimposed by a younger channel (**b**) that ends with a delta feature (**c**) in a nearby crater, which shows a small spill of and outflow into the older channel (**d**) indicate different periods of fluvial erosion and deposition (Mars Express HRSC images, ESA/DLR/FUB HSRC orbit 905, 8.65°N 48°W *scale bar* 5 km)

Fig. 13 Barchan dunes of fine grained dark material on Mars. (MGS MOC Image, MSSS. MOC FHA00451, 292.93°W 8.83°N, NASA/JPL/MSSS, MGS release no. MOC2-88, 11 March 1999)



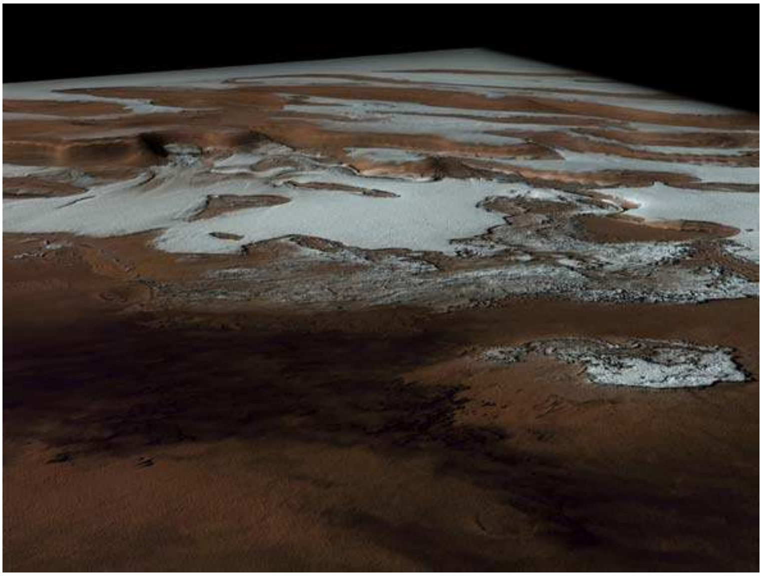


Fig. 14 Ice and layered terrain of the Martian North Pole. (Mars Express HRSC image ESA/DLR/FUB (G. Neukum) HSRC orbit 1167, 78°N 117°E, width of image in the front ∼100 km)

Apart from CO_2, the polar caps contain water ice about equivalent a global water layer about 0.6 m thick. In addition the polar layered terrains (Fig. 14) are assumed to contain the equivalent of a global water layer between 6–29 m deep (Kieffer and Zent 1992). The uppermost few meters of the Martian regolith contain about 35 ± 15 wt% H_2O, at latitudes $>60°$, as measured by the High Energy Neutron Detectors of the Mars Odyssey Mission (Feldman et al. 2002) and about 1 wt% in regions of lower latitude (Bieman et al. 1977; Boynton et al. 2002). The H_2O in the upper regolith is assumed to be not liquid but either pore ice, adsorbed water (Möhlmann 2004) or mineral bound water (Boynton et al. 2002). The total inventory of surface and near surface water is thus the equivalent of a global water layer about 7–30 m deep.

Clifford (1993) estimated the theoretical pore volume of the upper 8.5 km of the Martian crust at about 7.8×10^7 km^3 comparing to about 540 m equivalent global water layer. As the Martian geothermal heat flow decreases with time the thickenss of the permafrost (the cryosphere) increases, thus reducing the pore volume for liquid water. The thickness of the cryosphere is $z = \kappa[(T_\mathrm{m} - T_\mathrm{a})/Q]$, with κ = thermal conductivity, T_m = melting temperature of ground ice, T_a = mean annual temperature and Q = geothermal heat flux. The geothermal heat flux was about 150 mW/m^2 at the beginning and has decreased to about 30 mW/m^2 today (Spohn et al. 2001) resulting in a mean increase of the cryosphere (the permafrost layer) from about 0.5 km to over 3 km. If water is present in the cryosphere it is frozen. However, according to the theoretical pore volume of 7.8×10^7 km^3, between 5.7×10^7 km^3 and 2.9×10^7 km^3 should have been available over time for ground water beneath the cryosphere, i.e. equivalent to a global water layer of about 400 m to 200 m. Thus, the Martian subsurface has a high potential for water rich environments, either as ice in the cryosphere, or as liquid groundwater beneath it.

However, was there enough water available on Mars to fill the pore volume? There is much evidence for this on the surface of Mars, which exhibits fluvial and glacial erosion. The amount of water involved in this surface erosion has been estimated by a number of authors (e.g. Carr 1987, 1996; Clifford 1993; Baker et al. 1991; Baker 2001; Jaumann et al. 2002, 2005) using eroded volumes of material, erosion rates and water to sediment ratios. Although the different estimates vary between equivalent global water layer of 80 m to 1200 m, they suggest there should have been water available to fill at least part of the available pore volume.

Valley networks (Fig. 15), chaotic terrain and outflow channels (Fig. 16) are the surface expression of subsurface water release or precipitation episodes. However, almost

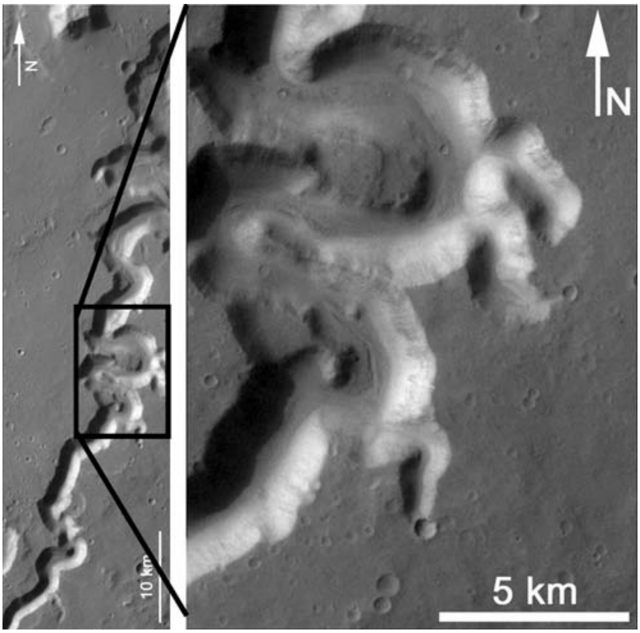


Fig. 15 Nanedi Vallis on Mars. Meander, terraces and interior channels indicate fluvial origin of the sinuous surface feature. (Mars Express HRSC image ESA/DLR/FUB (G. Neukum) HSRC orbit 905, 5.25°N 311.8°E)

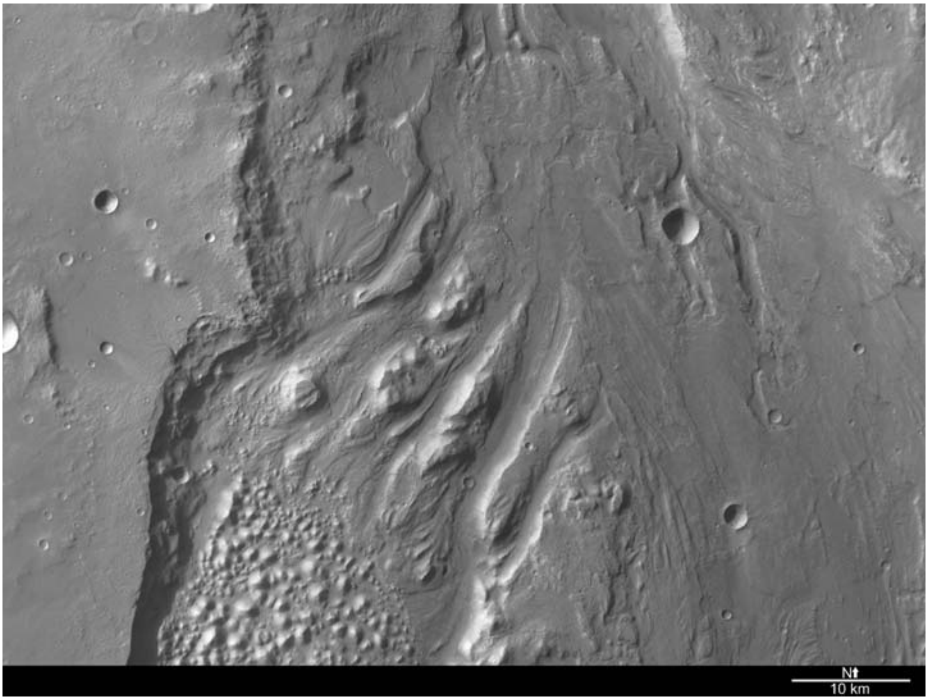


Fig. 16 Outflow channels in Iani Chaos on Mars. Large scale erosion and flow features indicate rapid water release. (Mars Express HRSC image ESA/DLR/FUB (G. Neukum) HSRC orbits 923 and 924, 1.5°N 342.2°E)

all of these features date back to 3.6 Ga or more ago (e.g. Scott and Dohm 1992; Carr 1995; Masson et al. 2001; Jaumann et al. 2002), indicating that significant surficial water environments outside the poles are very ancient and inactive today. Nevertheless, these fluvial features together with sedimentary features (e.g. Malin and Edgett 2000a; Squyres et al. 2004) (Fig. 17), show that water rich, and by implication, habitable environments on the surface existed in early Martian history.

However, all the fluvial and sedimentary features on Mars are relatively poorly developed, indicating inadequate drainage and erosion (Carr 1995; Carr and Malin 2000; Jaumann et al. 2005), and thus restricted time frames for water being present on the surface. This is consistent with the distribution of hydrated minerals (Bibring et al. 2005).

Relatively young erosion features (a few tens to hundreds of million years old) on Mars appear to be of glacial origin (Neukum et al. 2004; Hauber et al. 2005b; Head et al. 2005). Recent water related surface features are restricted to small scale gullies (Malin and Edgett 2000b) and wet debris flows (Reiss and Jaumann 2003). The gully morphologies are most consistent with a formation from near surface aquifers (Heldman and Mellon 2004). The discovery of very recently formed light toned gully deposits suggest, that liquid water flowed on the surface of Mars during the last decade (Malin et al. 2006).

In summary, although there are rare local recent water environments outside the polar regions, most of the water is expected in the porous subsurface and all traces of significant amounts of water on the surface are old (Fig. 18).

Fig. 17 Layered deposits in the Valles Marineris on Mars indicate sedimentation. The layers are composed of the sulfates gypsum and kieserite (Bibring et al. 2005). (Mars Express HRSC image ESA/DLR/FUB (G. Neukum) HSRC orbit 243, 4.4°S 297.7°E length of deposit 60 km)

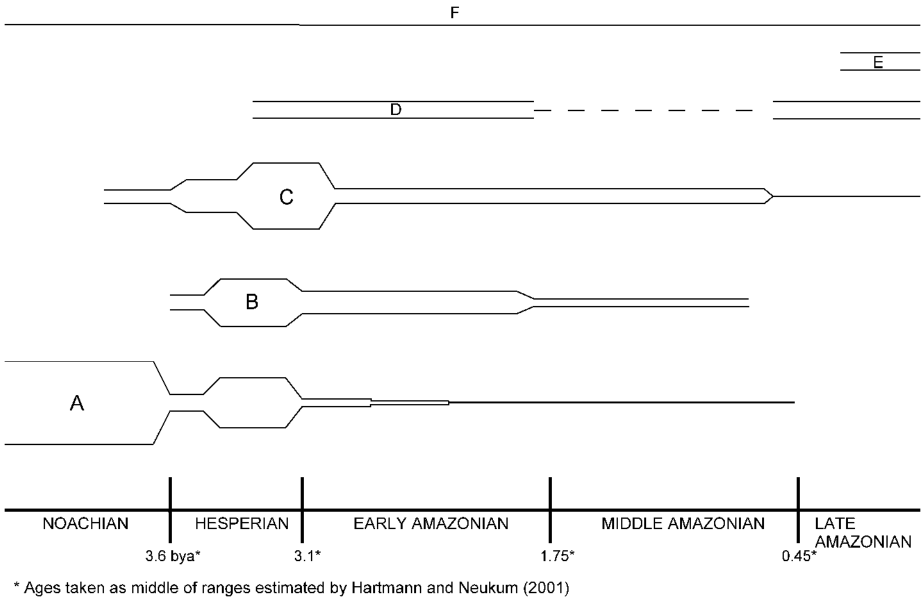


Fig. 18 Evolution of surface water and ice throughout the Martian history. *A*... valley networks; *B*... lacustrine deposits; *C*... outflow channels; *D*... glacial and periglacial deposits; *E*... polar caps, low latitude glaciers gullies; *F*... aeolian activity. (For references see e.g. Carr 1996; Malin and Edgett 2000a, 2000b; Masson et al. 2001; Jaumann et al. 2002; Head et al. 2003; Jaumann 2003; Head et al. 2005; Hauber et al. 2005b)

16 Crustal Geothermal Processes on Mars

Creating a habitable environment is a complex process involving a wide variety of interacting processes. But an energy source is the first prerequisite for any biological activity. The other prerequisite is temperature: life needs liquid. The seasonal average surface temperature is below freezing (though summer daytime temperatures can be above the freezing point

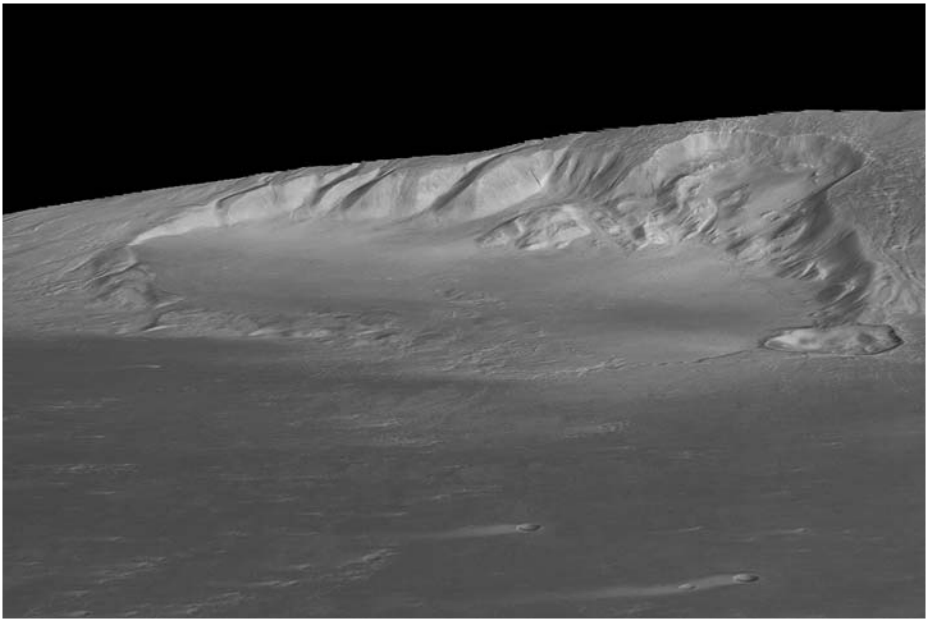


Fig. 19 Thermal modeling (Helbert et al. 2005) indicates that these glacial deposit on the northwestern flank of Hecates Tholus still contain ice (Hauber et al. 2005b). (Mars Express HRSC image ESA/DLR/FUB (G. Neukum) Location 148.8°E 32.8°N (*center of image*))

of liquid water, though not above the triple point). In the past and in the future, as orbital cycles produce periods of more intense local solar forcing, Mars has more clement episodes. Can life survive between periods of clement conditions by retreating underground?

Unfortunately our knowledge of geothermal processes or general thermal processes in the crust or even in the near surface of Mars is still very limited. Calculated geotherms (strictly termed Aresotherms, for Ares, Mars, as compared to Geos, Earth) from measurements of heat producing elements in Martian samples show that the interior is cold for many kilometers down. Cold, of course, is a subjective term: Hoffman (2001) estimated a crustal temperature increase with depth of 6.4–10.6 K/km.

Volcanoes provide activity and habitat. There is ample morphological evidence for continuous but episodic volcanic activity over the geological history of Mars—especially now with the image dataset obtained by Mars Express (for example Neukum et al. 2004). The youngest ages determined by crater size-frequency measurements are about 2 Ma ago, suggesting that the volcanoes are potentially still active today (Werner 2005). While we have not yet observed any modern volcanic activity it seems that the likelihood for localized hot spot activity or hydrothermal systems is high.

New infrared spectrometers (TES on Mars Global Surveyor, THEMIS on Mars Odyssey and OMEGA and PFS on Mars Express) have permitted a more detailed insight into the thermophysical properties of the surface of Mars (Christensen et al. 2004; Putzig et al. 2005; Formisano et al. 2005). Our knowledge of these properties is however still limited to the first few millimeters of the surface. Everything below this is accessible only by modeling. The stability of subsurface ice and the possibility of extant stagnant ice cores within glacial deposits remain controversial. It has been shown recently using thermophysical modeling (Helbert et al. 2005) that a morphologically identified glacial deposit on the northwestern

Fig. 20 A steaming fumarole on Mount Erebus with a 10 m ice-tower precipitated from H_2O-rich volcanic gas (from Hoffman and Kyle 2003)

Fig. 21 Inside a grotto below an ice-tower (from Hoffman and Kyle 2003)

flanks of Hecates Tholus (Fig. 19) (Hauber et al. 2005b) might still contain a stagnant ice core. Head et al. (2005) report several units on Mars which morphological evidence suggests may be glacial deposits. Many of these units are found on the flanks of volcanic edifices.

Hoffman and Kyle (2003) suggested the ice towers of Mt. Erebus (Figs. 20, 21) as analogues of biological refuges on Mars. They combined the idea of still existing near surface ice deposits with the assumption that there is still some localized volcanic activity on Mars today.

There are several examples from Mars that show a direct interaction between lava and ice in the geological history of Mars. The most obvious cases are the rootless cones seen in the northern lowlands. HRSC images show direct and violent interaction in the relatively recent geological history, for example at the scarps of Olympus Mons. Mars today is in relatively dormant phase, and any interactions which might be occurring today are presumably on a much less dynamic scale. Nevertheless, they may be driving local hydrothermal systems. Studying the geothermal processes in the first few tens to hundreds of meters below the surface of Mars today might thus uncover a wide variety of new habitats where biological activity may survive on this cold and dry planet.

17 The Evidence that Habitability Was Achieved: Early Life and Its Signatures

We know the Earth was habitable, but if we are to seek out the habitability of other planets, we must ask "what evidence do we have on Earth that habitability was actually achieved?". It is this type of evidence that we must use as a guide in seeking out life elsewhere.

Unfortunately the very dynamism of the Earth, essential for the upkeep of its habitability, is also the cause of the destruction of the earliest history of life. Plate tectonic recycling of the earliest continents and oceanic crust has effectively erased almost the first billion years of rock record. Although ancient zircons dating back to about 4.4. Ga exist (Wilde et al. 2001), and although ancient gneisses from about 4.0 Ga are exposed in Canada (Bowring and Housh 1995) and 3.7–3.8 Ga) metasediments occur at Isua, Greenland, the oldest low metamorphic grade sedimentary rocks are only ∼3.5 Ga old. Sedimentary rocks are important repositories of life's history because of their formation in aqueous environments: as we have seen above, water is the "spring of life" (Brack 2002).

It is not the aim of this chapter to air the heated debates surrounding the oldest traces of life (see review in Westall and Southam 2006). Suffice it to say that volcanic and chemical sediments in the ancient terrains (∼3.5–3.3 Ga) of the Pilbara in NW Australia and Barberton in South Africa contain a plethora of traces pertaining to life—carbon and nitrogen isotopic signatures, as well as macroscopic to microscopic traces in the form of stromatolites, microbial mats, biofilms and colonies of microorganisms. It is interesting to relate the occurrence of these traces to the habitability concept (Fig. 22).

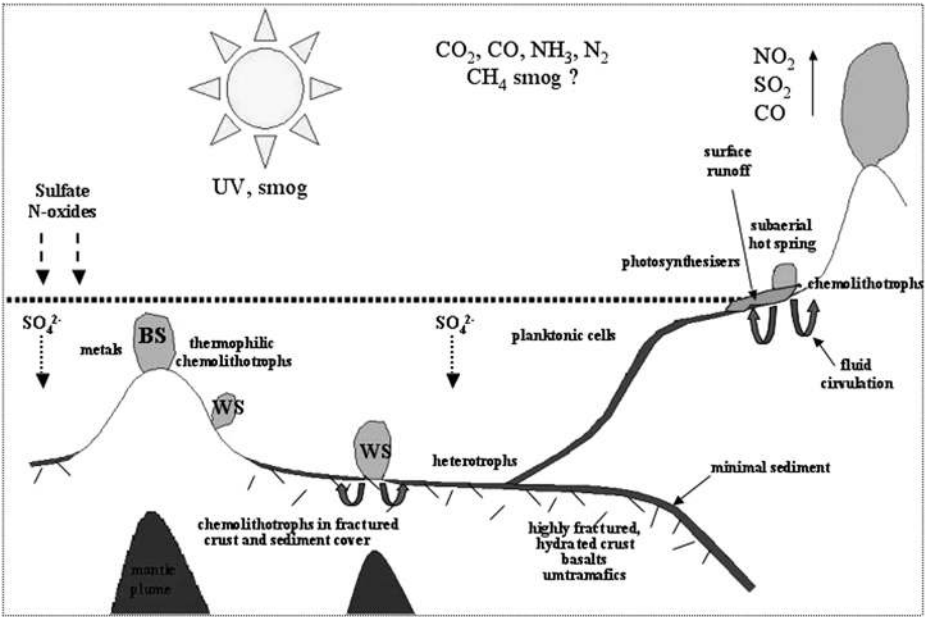


Fig. 22 Sketch showing the potential microbial habitats in the early-Mid Archean period. *BS*—black smoker; *WS*—white smoker. In the deep sea area, the potential habitats include the fractured, hydrated crust, the thin layer of sediments covering the crust, the black and white smokers, and the water column. In the shallow water regions around the emerged land masses (mainly volcano tops and emerged platforms), again sediment and fractured crust provide potential habitats, but the surfaces (subaqueous and subaerial sediment, lava, mineral) exposed to the sunlight represent habitats for potential photosynthesizers. Sketch based on Nisbet and Sleep (2001) and Westall and Southam (2006)

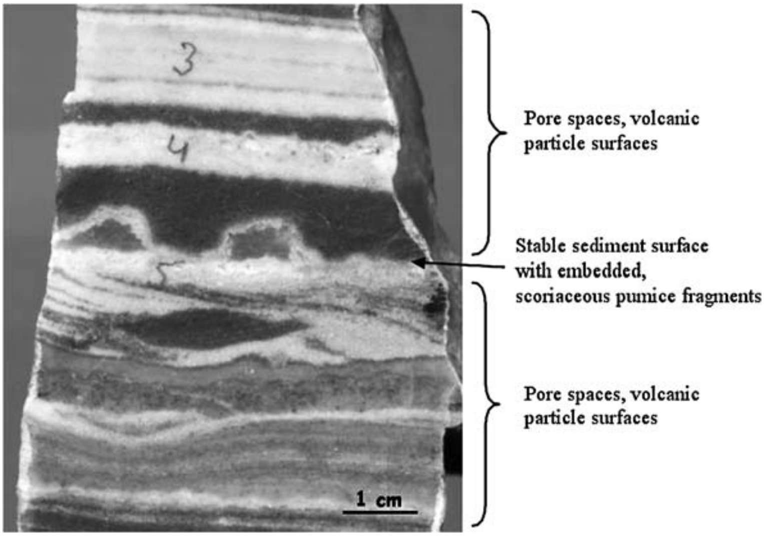


Fig. 23 Microhabitats in shallow water, water-logged coastal, volcanic sediments from the "Kitty's Gap Chert", Pilbara, W. Australia (3.446 Ga). Two main microhabitats are shown—the pore spaces and particle surfaces in the subsurface (ideal habitat for chemolithotrophic microganisms) and a sediment surface that was in the sunlight zone and stable for a short period of time (habitat for mat-building photosynthesisers)

We have seen above that the microhabitats available to life on early Earth were varied: the most protected being subsurface niches, sheltered from UV radiation and the occasional impact (cf. Lowe et al. 2003): the flux of UVb and UVc (DNA-damaging ultra-violet, the radiation wavelengths most damaging to the genetic machinery of organisms), is estimated to have been up to 1,000 times greater than today (Cockell 2000).

Clear evidence of life based apparently on a chemolithotrophic metabolism (whereby inorganic sources of carbon and energy are used) is to be found in the waterlogged, subsurface sediments of volcanic sands and silts deposited in a mudflat environment on the edges of a shallow water basin reported by De Vries (2004); Westall et al. (2006a); and Orberger et al. (2006) (Figs. 23, 24). These sediments host colonies of coccoidal microorganisms on the surfaces of the volcanic grains. Both shallow and deeper water marine sediments contain the resedimented remains of planktonic prokaryotes and shallow water microbial mats (Walsh 1992; Tice and Lowe 2004; Westall et al. 2006aa). Hydrothermal chert veins contain clots and wisps of carbonaceous material of enigmatic origin but with isotopic carbon and nitrogen signatures that are consistent with microbial fractionation (Ueno et al. 2004; Pinti and Hashizume 2001; Pinti et al. 2001; Orberger et al. 2006). It has been suggested that this type of carbon is a purely abiogenic Fischer–Tropf synthesis (van Zuilen et al. 2002, 2005; see Westall and Southam 2006 for a discussion). Other putative traces of life in a subsurface environment, namely corrosion pits in the vitreous surfaces of pillow lavas (Furnes et al. 2004), have been evoked above. The possibility of microbial involvement in the formation of "stalactites" in chert-filled vadose zone cracks in these Early Archean sediments is also under investigation (Hofmann et al. 2006).

Despite the supposedly hostile conditions reigning at the surface of the Earth in the Archean (high UVb and UVc radiation and impact extinction), there are numerous indications that life was not particularly perturbed and that biological activity not only survived but flourished. Apart from the few resedimented remains of shallow water microbial mats,

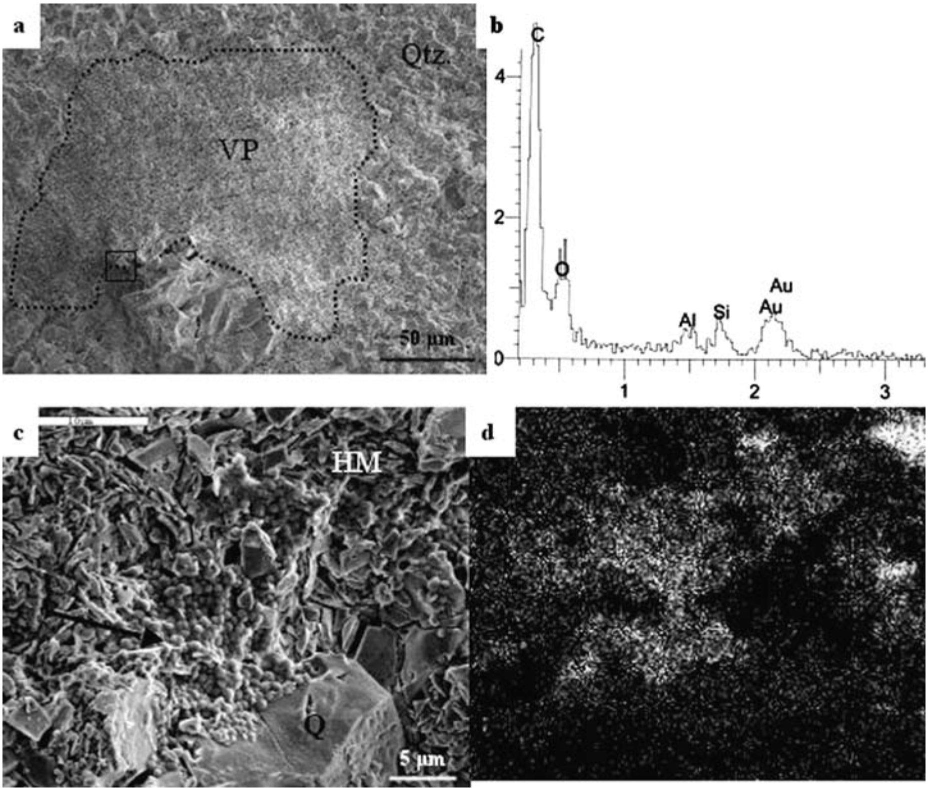


Fig. 24 Silicified microbial colonies in the "Kitty's Gap Chert" sediments. A volcanic particle (**a**) hosts colonies (**c**) of carbonaceous (**b**) coccoidal microorganisms (**d**) (Westall et al. 2006aa)

the silicified remains of mid- to late Archean in situ mats are relatively common (Walsh 1992, 2004; Westall et al. 2006b).

In general, all that remains of the mats are compacted layers of carbonaceous-rich horizons in which occasionally stray filaments or bundles of filaments are trapped in a silica matrix. Only where rapid fossilization caused by the deposition of a protective layer of silica, as well as silica impregnation, has occurred, are these mats preserved well enough to provide further information useful for environmental interpretations. Such a case is a filamentous mat that formed 3.33 Ga ago on a beach surface under evaporitic conditions (Westall et al. 2006b, 2006c) (Fig. 25). Given the apparently healthy appearance of the microorganisms prior to their silicification (from an outflow of hydrothermal silica that effectively killed the mat off), it is clear that they were well adapted for coping with adverse environmental conditions.

Another interesting characteristic of this and other microbial mats from this period is that they were most likely formed by anoxygenic photosynthesisers (Westall 2003; Walsh 2004; Tice and Lowe 2004; Westall et al. 2006b); indeed, mat-forming behaviour is, today, typical of photosynthetic microorganisms. Mats readily form even in ephemeral environments (Stahl 1994). The silicified Early Archean beach mat described above is just one such ephemeral environment. These mats are, however, poorly developed in the sense that, even though anoxygenic photosynthetic metabolisms provide far more energy and therefore al-

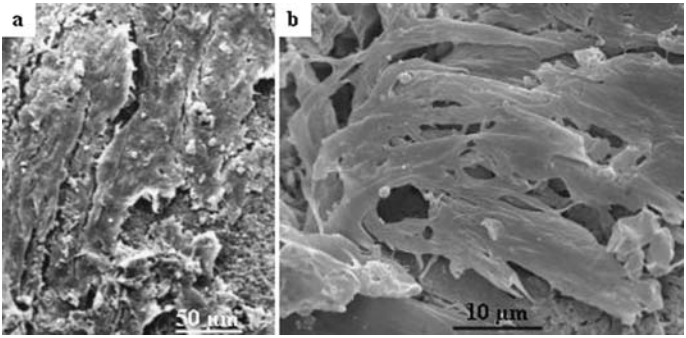


Fig. 25 Silicified microbial mat formed on a beach surface under evaporitic conditions (Westall et al. 2006b). (**a**) Plan view of the evaporite mineral-encrusted and desiccated filamentous mat. (**b**) Portions of the mat exhibit overturned filaments and bundles of filaments indicating initial formation under flowing water

low greater biological productivity than chemolithotrophic metabolisms, it still takes a long time for a mat to be developed.

In comparison, the meter-high stromatolites found today in the protected habitat of Shark Bay, Western Australia, formed by highly efficient oxygenic photosynthesising primary producers (plus associates), are estimated to be several thousands of years old. This means that, for any serious microbial mat development, the environmental conditions need to be stable over long periods of time. The Early-Mid Archean environment, with its very active tectonic cycle and rapidly changing basin edge scenarios, was not particularly conducive to the construction of layers upon layers of slowly accumulating microbial mats.

There are some exceptions however and they are important also for the implications regarding the changing tectonic regime of the geologically evolving Earth. Small stromatolites do occur in the Pilbara and in Barberton (Byerly et al. 1986; Hofmann et al. 1999; Allwood et al. 2006). Averaging about 10 cm or less in height, these bio-constructions are far smaller than their larger, Later Archean to recent counterparts that testify to the effectiveness of oxygenic photosynthesis. Despite their small size, the Early-Mid Archean stromatolites probably required equally as long to grow, given the limitations of anoxygenic photosynthesis. However, the important implication is that, at least in certain small areas of the Earth at this period, stabilized continents with stable, shallow water continental platforms were starting to come into existence. It was only through this geological evolution that suitable habitats started to appear that provided the breeding grounds for opportunistic anoxygenic photosynthesiers first and then the oxygenic photosynthesisers that made our planet what it is today.

18 Sustaining Habitability: The Long Term Scale—Is There an Outlook?

Is Mars a lesson? Does habitability end? Will Earth become Mars-like?—or, for that matter, Venus-like? Why has Earth, uniquely, sustained habitability?

The habitable zone is usually defined as the range of distances from the Sun where liquid water is stable on the planet's surface. This definition has both parochial and operational aspects. The latter in particular focus attention on the kinds of extrasolar planets we might plausibly detect within the horizons of our lives. The conventional definition excludes planets or moons for which liquid water is only present in clouds, or only present underground or under ice cover. It is widely thought that life on such extrasolar planets or moons would be undetectable to telescopic inspection of extrasolar planets; in any event, if such life is present now in our own Solar System, it has remained hidden from us.

The inner and outer limits of the habitable zone are determined by the evaporation and freezing of water, respectively. Neither has been as clearly defined as we should like. A hard inner boundary is set by the runaway greenhouse limit (see chapter by Zahnle et al.). If the planet were any closer to the Sun, all liquid water on the surface would evaporate. There is some uncertainty in the precise location of the inner bound that depends on how much sunlight the planet reflects, compared to how much it absorbs. For a cloudy ocean world the reflectivity would probably be around 40%.

Kasting et al. (1993) defined the inner limit to habitability as the threshold for hydrogen escape. This is, however, not necessarily the crucial limiting factor. On Earth today hydrogen escape is tiny because the top of the troposphere is cold enough that very little water reaches the stratosphere or above. With very little hydrogen at high altitudes there is very little hydrogen escape. If the Earth were somewhat closer to the Sun (0.95 AU) this atmospheric cold trap would disappear and the limit on hydrogen escape would not exist. Kasting et al. called this the "moist greenhouse". However, such a planet would have liquid water oceans—possibly for a very long time albeit not forever. Thus the hydrogen-threshold is not obviously the limiting factor, especially if life can appear and evolve complexity faster than it has done on Earth—say in a few hundred million years.

A second issue arises in the difference between wet and dry planets. The runaway greenhouse limit applies only if water condenses in the atmosphere or at the surface. If a planet is dry enough its tropical atmosphere can be truly cloud-free. If so, it can be sited closer to the Sun than a wet planet and still have condensed water near the poles (Abe and Abe-Ouchi 2005). Because hydrogen escapes before the ocean evaporates, Kasting et al.'s (1993) moist greenhouse could evolve into a dry planet without ever becoming uninhabitable. Indeed, it is an interesting speculation if Venus could remain habitable by aerophile acidophiles floating in the upper atmospheric smog.

The outer limit to the habitable zone is determined by the strength of the greenhouse effect. In principle there isn't an outer limit because, given the right gases and the will to use them, there is no upper limit on the greenhouse effect. Stevenson (1999) pointed out that a thick hydrogen atmosphere could even make a planet lost in space habitable. But would such a planet be recognizable as habitable? For specificity Kasting et al. (1993) took the conservative view that CO_2 is responsible for the greenhouse effect. CO_2 condenzation then sets the outer boundary at $\sim$1.7 AU today in our Solar System.

How long can an Earth-like planet remain habitable? Figure 27 explores the bounds of habitability for an Earth-like planet at a given current distance. Habitable planets have been defined as receiving an insolation S between 0.7 and 1.4 times what Earth receives today. These bounds correspond to the bounds for the habitable zone as defined in Fig. 26. The more restrictive habitable zone for insolation $0.75 < S < 1.1$ corresponds to the habitable zone originally defined by Kasting et al. (1993); the inner bound the "moist" greenhouse of the other figure. The less restrictive $0.4 < S < 1.4$ sets the inner boundary by the runaway greenhouse effect and the outer boundary includes a rough upper limit on the greenhouse warming that can be obtained from CO_2 clouds (Kasting 1996, 1997). The figure was prepared using standard models of solar evolution applied over the full lifetime of our Sun. Late in the Sun's life, when it is a red giant, it becomes much brighter than it is now. During this time the habitable zone leaves the realm of the terrestrial planets, but extends into the realm of the giant planets. The structure in the curve arises from stationary points in the Sun's evolution.

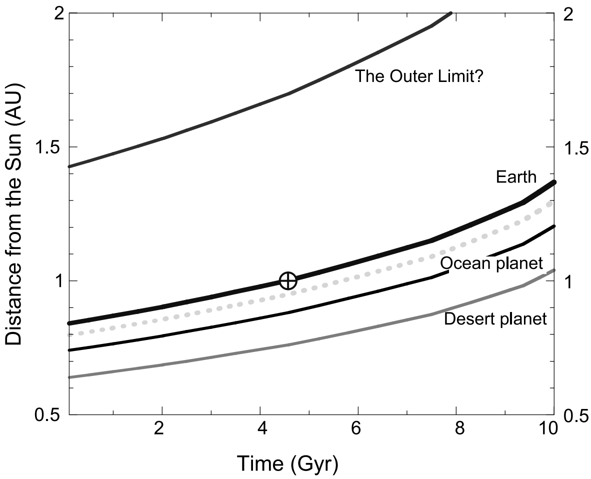


Fig. 26 The habitable zone over the first ten Ga of solar evolution. Here we consider the long-term habitability of an Earth-like planet as the Sun evolves. Solar evolution is from Sackmann et al. (1993). The inner edge of the habitable zone is set by the runaway greenhouse effect. This is more permissive for a desert planet (like Mars) than it is for a planet with oceans of water (like Earth). The less permissive *dashed curve* refers to the "moist" greenhouse, which Kasting et al. (1993) identify with the inner edge of the habitable zone. The moist greenhouse describes a perfectly habitable ocean planet that suffers rapid hydrogen escape; the importance of this is debatable. The outer edge of the habitable zone as drawn here is set by the condenzation of CO_2 clouds (Kasting et al. 1993). This is conservative. If the CO_2 clouds themselves have a strong greenhouse warming effect (Forget and Pierrehumbert 1997), or if other less volatile greenhouse gases are abundant, the habitable zone expands. Stevenson (1999) argued that, because there is no upper bound on the greenhouse effect, there is no outer limit to the habitable zone

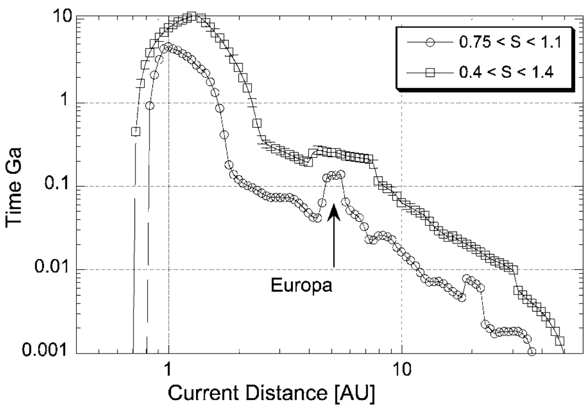


Fig. 27 The wave of quickening. As the Sun evolves, the habitable zone sweeps through the solar system (assuming the standard solar evolution from Sackmann et al. 1993). Figure 27 shows how long any location in our Solar System remains in the habitable zone. More and less generous assessments of the habitable zone are shown. It is perhaps no accident that Earth-bound natural philosophers find that Earth is in the best place in the Solar system

The main story is one of steady brightening during the roughly 11 Ga that the Sun spends on the main sequence. This is the longest stage in the Sun's life, when it fuses hydrogen into helium in its core. The core is only 10% of the Sun's mass. When hydrogen in the core is exhausted, the Sun begins to fuse the rest of its hydrogen into helium, but to do so the Sun must get hotter inside and therefore gets brighter. Thus begins the red giant phase, which lasts about 1 Ga and consumes much of the remaining hydrogen. As it runs out of hydrogen, the Sun becomes very big and bright and the habitable zone sweeps to Uranus and beyond. At a certain point the helium in the core begins to fuse into carbon. When this happens the

Sun enters a 100 million year stage of relative constancy that happens to put Jupiter and its moons in the habitable zone. This is the reason why there is a local maximum, marked "Europa", in Fig. 27.

There, perhaps, lies our last home.

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References

Y. Abe, A. Abe-Ouchi, Stability of liquid water on a land planet: Wider habitable zone for a less water planet than an aqua planet. American Geophysical Union, Fall Meeting 2005, abstract #P51D-0965, 2005

C.M. Aitken, D.M. Jones, S.R. Larter, Nature **431**, 291–294 (2004)

A.C. Allwood, M.R. Walter, B.S. Kamber, C.P. Marshall, I.W. Burch, Nature **441**, 714–718 (2006)

R.E. Arvidson, E.A. Guinness, S. Lee, Nature **278**, 533–535 (1979)

R.E. Arvidson, F. Poulet, J.-P. Bibring, M. Wolff, A. Gendrin, R.V. Morris, J.J. Freeman, Y. Langevin, N. Mangold, G. Bellucci, Science **307**, 1591–1593 (2005)

V.R. Baker, R.G. Strom, S.K. Croft, V.C. Gulick, J.S. Kargel, G. Komatsu, Nature **352**, 589–594 (1991)

V.A. Baker, M.H. Carr, V.C. Gulick, C.R. Williams, M.S. Marley, in *Mars*, ed. by H.H. Kieffer, B.M. Jakosky, C.W. Snyder, M.S. Matthews (University of Arizona Press, Tucson, 1992), pp. 493–522.

V.A. Baker, Nature **412**, 228–236 (2001)

L.G. Benning, V.R. Phoenix, N. Yee, M.J. Tobin, Geochim. Cosmochim. Acta **68**, 729–741 (2004)

K. Benzerara, N. Menguy, F. Guyot, F. Skouri, G. de Lucca, T. Heulin, Earth Planet. Sci. Lett. **228**, 439–449 (2004a)

K. Benzerara, T.-H. Yoon, T. Tyliszczak, B. Constantz, A.M. Spormann, G.E. Brown Jr., Geobiology **2**, 249–259 (2004b)

K. Benzerara, T.-H. Yoon, N. Menguy, T. Tyliszczak, G.E. Brown Jr., Proc. Natl. Acad. Sci. USA **102**, 979–982 (2005)

J.-P. Bibring, Y. Langevin, A. Gendrin, B. Gondet, F. Poulet, M. Berthé, A. Soufflot, R. Arvidson, Mangold, J. Mustard, P. Drossart, the OMEGA team, Science **307**, 1576–1581 (2005)

H. Bieman, J. Oró, P. Toulmin, L.E. Orgel III, A.O. Nier, D.M. Anderson, P.G. Simmonds, D. Flory, A.V. Diaz, D.R. Rushneck, J.E. Biller, A.L. LaFleur, J. Geophys. Res. **82**, 4641–4658 (1977)

P.J. Boston, M.N. Spilde, D.E. Northup, L.A. Melim, D.S. Soroka, L.G. Kleina, K.H. Lavoie, L.D. Hose, L.M. Mallory, C.N. Dahm, L.J. Crossey, R.T. Schelble, Astrobiology **1**, 25–55 (2001)

W.V. Boynton, W.C. Feldman, S.W. Squyres, T. Prettyman, J. Brückner, L.G. Evans, R.C. Reedy, R. Starr, J.R. Arnold, D.M. Drake, P.J.A. Englert, A.E. Metzger, I. Mitrofanov, J.I. Trombka, C. d'Uston, H. Wänke, O. Gasnault, D.K. Hamara, D.M. Janes, R.L. Marcialis, S. Maurice, I. Mikheeva, R. Torkar, C. Shinohara, Science **297**, 81–85 (2002)

S.A. Bowring, T. Housh, Science **269**, 1535–1540 (1995)

A. Brack, in *Astrobiology: the Quest for the Conditions of Life*, ed. by G. Horneck, C. Baumstark-Khan (Springer, Berlin, 2002), pp. 79–88.

M.D. Brasier, O.R. Green, J.F. Lindsay, N. McLoughlin, A. Steele, C. Stoakes, Precambr. Res. **140**, 55–102 (2005)

D. Breuer, T. Spohn, J. Geophys. Res. **108**, E/ (2003). doi:10.1029/2002JE001999

J.C. Bridges, P.H. Warren, J. Geol. Soc. London **163**(2), 229–251 (2006)

D.M. Burr, J.A. Grier, A.S. Mcewen, L.P. Keszthelyi, Icarus **159**, 53–73 (2002)

G.R. Byerly, M.M. Walsh, D.L. Lowe, Nature **319**, 489–491 (1986)

D.E. Caldwell, S.J. Caldwell, Microb. Ecol. **47**, 252–265 (2004)

R.A. Carddock, T.A. Maxwell, J. Geophys. Res. **98**, 3453–3468 (1993)

M.H. Carr, Nature **326**, 30–35 (1987)

M.H. Carr, H. Wänke, Icarus **98**, 61–71 (1992)

M.H. Carr, J. Geophys. Res. **100**, 7479–7507 (1995)

M.H. Carr, *Water on Mars* (Oxford University Press, 1996), 229 pp.

M.H. Carr, M.C. Malin, Icarus **146**, 366–386 (2000)

S. Castanier, G. Le Metayer-Levrel, J.P. Perthuisot, Sediment. Geol. **126**, 9–23 (1999)

D.C. Catling, K.J. Zahnle, C.P. McKay, Science **293**, 839–843 (2001)

C.S. Chan, G. De Stasio, S.A. Welch, M. Girasole, B.H. Frazer, M.V. Nesterova, S. Fakra, J.F. Banfield, Science **303**, 1656–1658 (2004)
F.H. Chapelle, K. O'neill, P.M. Bradley, B.A. Methé, S.A. Ciufo, L.L. Knobel, D.R. Lovley, Nature **415**, 312–315 (2002)
A.E. Chichibabin, *Basic Principals of Organic Chemistry*, vol. 1 (Goskhimizdat, Leningrad, 1953) 796 pp. (in Russian)
P.R. Christensen, B.M. Jakosky, H.H. Kieffer, M.C. Malin, H.Y. McSween Jr., K. Nealson, G.L. Mehall, S.H. Silverman, S. Ferry, M. Caplinger, M. Ravine, Space Sci. Rev. **110**, 85–130 (2004)
S.M. Clifford, J. Geophys. Res. **98**, 10973–11016 (1993)
S.M. Clifford, T.J. Parker, Icarus **154**, 40–79 (2001)
C.S. Cockell, Planet. Space Sci. **4**, 203–214 (2000)
J.E.P. Connerney, M.H. Acuña, P. Wasilevski, N.F. Ness, H. Rèmw, C. Mazelle, D. Vignes, R.P. Lin, D. Mitchell, P. Cloutier, Science **284**, 794–798 (1999)
J.E.P. Connerney, M.-H. Acuna, N.F. Ness, G. Kletetschka, D.L. Mitchell, R.P. Lin, H. Reme, PNAS **102**(42), 14970–14975 (2005)
J.C. Dann, A.H. Holzheid, T.L. Grove, H.Y. McSween, Meteorit. Planet. Sci. **36**, 793–806 (2001)
S.T. De Vries, Early Archean sedimentary basins: depositional environment and hydrothermal systems. Examples from the Barberton and Coppin Gap greenstone belts. Geologica Ultraiectina, University of Utrecht, 2004, 159 pp.
S. D'Hondt, B.B. Jørgensen, J. Miller, A. Batzke, R. Blake, B.A. Cragg, H. Cypionka, G.R. Dickens, T. Ferdelman, K.-U. Hinrichs, N. G. Holm, R. Mitterer, A. Spivack, G. Wang, B. Bekins, B. Engelen, K. Ford, G. Gettemy, S.D. Rutherford, H. Sass, C.G. Skilbeck, I.W. Aiello, G. Gue'Rin, C.H. House, F. Inagaki, P. Meister, T. Naehr, S. Niitsuma, R.J. Parkes, A. Schippers, D.C. Smith, A. Teske, J. Wiegel, C. Naranjo Padilla, J.L. Solis Acosta, Science **306**, 2216–2221 (2004)
S. D'Hondt, S. Rutherford, A.J. Spivack, Science **295**, 2067–2070 (2002)
G. Dreibus, H. Wänke, Icarus **71**, 225–240 (1987)
G. Dreibus, H. Wänke, in *Origin and Evolution of Planetary and Satellite Atmospheres*, ed. by S.K. Atria, J.B. Pollack, M.S. Matthews (Univ. of Arizona Press, Tucson, 1989), pp. 268–288
D. Emerson, W.C. Ghiorse, Appl. Env. Microbiol. **58**, 4001–4010 (1992)
B. Fegley Jr., R.G. Prinn, H. Hartman, G.H. Watkins, Nature **319**(6051), 305–308 (1986)
W.C. Feldman, W.V. Boynton, R.L. Torkar, T.H. Prettyman, O. Gasnault, S.W. Squyres, R.C. Elphic, D.J. Lawrence, S.L. Lawson, S. Maurice, G.W. McKinney, K.R. Moore, R.C. Reedy, Science **297**, 75–78 (2002)
F. Forget, R. Pierrehumbert, Science **278**, 1273–1276 (1997)
V. Formisano, S. Atreya, T. Encrenaz, N. Ignatiev, M. Giuranna, Science **306**, 1758–1761 (2004)
V. Formisano, F. Angrilli, G. Arnold, S. Atreya, G. Bianchini, D. Biondi, A. Blanco, M.I. Blecka, A. Coradini, L. Colangeli, A. Ekonomov, F. Esposito, S. Fonti, M. Giuranna, D. Grassi, V. Gnedykh, A. Grigoriev, G. Hansen, H. Hirsh, I. Khatuntsev, A. Kiselev, N. Ignatiev, A. Jurewicz, E. Lellouch, J. Lopez Moreno, A. Marten, A. Mattana, A. Maturilli, E. Mencarelli, M. Michalska, V. Moroz, B. Moshkin, F. Nespoli, Y. Nikolsky, R. Orfei, P. Orleanski, V. Orofino, E. Palomba, D. Patsaev, G. Piccioni, M. Rataj, R. Rodrigo, J. Rodriguez, M. Rossi, B. Saggin, D. Titov, L. Zasova, Planet. Space Sci. **53**(10), 963–974 (2005)
R.B. Frankel, D.A. Bazylinski, Rev. Mineral. Geochem. **54**, 95–114 (2003)
H.V. Frey, J.H. Roark, K.M. Shockey, E.L. Frey, S.E.H. Sakimoto, Geophys. Res. Lett. **29**(10), 1384 (2002). doi:10.1029/2001.GL013832
H. Furnes, K. Muehlenbachs, O. Tumyr, T. Torsvik, I.H. Thorseth, Terra Nova **11**, 228–233 (1999)
H. Furnes, H. Staudigel, Earth Planet. Sci. Lett. **166**, 97–103 (1999)
H. Furnes, N.R. Banerjee, K. Muehlenbachs, H. Staudigel, M. de Wit, Science **304**, 578–581 (2004)
H. Furnes, N.R. Banerjee, K. Muehlenbachs, A. Kontinen, Precambr. Res. **136**, 125–137 (2005)
S.J.G. Galer, K. Mezger, Precambr. Res. **92**, 389–412 (1998)
R. Gellert, R. Rieder, R.C. Anderson, J. Brückner, B.C. Clark, G. Dreibus, T. Economou, G. Klingelhöfer, G.W. Lugmair, D.W. Ming, S.W. Squyres, C. d'Uston, H. Wänke, A. Yen, J. Zipfel, Science **305**, 829–832 (2004)
A. Gendrin, N. Mangold, J.-P. Bibring, Y. Langevin, B. Gondet, F. Poulet, G. Bonello, C. Quantin, J. Mustard, R. Arvidson, S. LeMouélic, Science **307**, 1587–1590 (2005)
M.V. Gerasimov, in *Catastrophic Events and Mass Extinctions: Impacts and Beyond*, ed. by C. Koeberl, K.G. MacLeod. Special Paper No 356 (Geological Society of America, Boulder, 2002), pp. 705–716
M.V. Gerasimov, L.M. Mukhin, M.D. Nussinov, Doklady Akademii Nauk of the USSR **275**(3), 646–650 (1984)
M.V. Gerasimov, Yu.P. Dikov, O.I. Yakovlev, F. Wlotzka, in *Lunar and Planetary Science XXV, Lunar and Planet. Inst.* (Abstracts) (Houston, 1994a), pp. 413–414

M.V. Gerasimov, Yu.P. Dikov, O.I. Yakovlev, F. Wlotzka, in *Lunar and Planetary Science XXV, Lunar and Planet. Inst.* (Abstracts) (Houston, 1994b), pp. 415–416
M.V. Gerasimov, Yu.P. Dikov, O.I. Yakovlev, F. Wlotzka, in *Large Meteorite Impacts and Planetary Evolution* (abstract), Sudbery, LPI Contr. No 922, 1997, pp. 15–16
M.V. Gerasimov, B.A. Ivanov, O.I. Yakovlev, Yu.P. Dikov, in *Laboratory Astrophysics and Space Research*, ed. by P. Ehrenfreund, K. Krafft, H. Kochan, V. Pirronello. Astrophysics and Space Science Library, vol. 236 (Kluwer, Dortrecht, 1999), pp. 279–330
M.V. Gerasimov, Yu.P. Dikov, O.I. Yakovlev, F. Wlotzka, Deep-sea research Part II. Top. Stud. Oceanogr. **49**(6) 995–1009 (2002)
W.C. Ghiorse, Annu. Rev. Microbiol. **38**, 515–550 (1984)
T.M. Gihring, P.L. Bond, S. Peters, J.F. Banfield, Extremophiles **7**, 123–130 (2003)
M.P. Golombek, N.T. Bridges, J. Geophys. Res. **105**, 1841–1853 (2000)
N.V. Grassineau, P. Abell, P.W.U. Appell, D. Lowry, E.G. Nisbet, in *Evolution of Early Earth's Atmosphere, Hydrosphere and Biosphere – Constraints from Ore Deposits*, ed. by S.E. Kesler, H. Ohmoto (Geological Society of America, 2006), pp. 33–52
L.A. Haskin, A. Wang, B.L. Jolliff, H.Y. McSween, B.C. Clark, D.J. DesMarais, S.M. McLennan, N.J. Tosca, J.A. Hurowitz, J.D. Farmer, A. Yen, S.W. Squyres, R.E. Arvidson, G. Klingelhöfer, C. Schröder, P.A. de Souza Jr., D.W. Ming, R. Gellert, J. Zipfel, J. Brückner, J.F. Bell III, K. Herkenhoff, P.R. Christensen, S. Ruff, D. Blaney, S. Gorevan, N.A. Cabrol, L. Crumpler, J. Grant, L. Soderblom, Nature **436**, 66–69 (2005)
E. Hauber, K. Gwinner, D. Reiss, F. Scholten, G.G. Michael, R. Jaumann, G.G. Ori, L. Marinangeli, G. Neukum, The HRSC Co-Investigator Team, Lunar Planet. Sci. **36** (2005a), abstract # 1661
E. Hauber, S. van Gasselt, B. Ivanov, S.C. Werner, J.W. Head, G. Neukum, R. Jaumann, R. Greeley, M. Mitchell, P. Muller, the HRSC Co-Investigator Team, Nature **434**, 356–360 (2005b)
J.M. Hayes, J.R. Waldbauer, Philos. Trans. Roy. Soc. Lond. B Biol. Sci. **361**, 931–950 (2006)
J.W. Head, R. Greeley, M.P. Golombek, W.K. Hartmann, E. Hauber, R. Jaumann, P. Masson, G. Neukum, L.E. Nyquist, M.H. Carr, in *Chronology and Evolution on Mars*, ed. by R. Kallenbach, J. Geiss, W.K. Hartmann. Space Science Series of ISSI. Space Sci. Rev. **96**, 263–292 (2001)
J.W. Head, J.F. Mustard, M.A. Kreslavsky, R.E. Milliken, D.R. Marchant, Nature **426**, 797–802 (2003)
J.W. Head, G. Neukum, R. Jaumann, H. Hiesinger, E. Hauber, M. Carr, N. Mangold, M. Kreslavsky, S. Milkovich, the HRSC Co-Investigator Team, Nature **434**, 346–351 (2005)
J. Helbert, D. Reiss, E. Hauber, J. Benkhoff, Geophys. Res. Lett. **32**(17), L17201 (2005). CiteID
J.L. Heldman, M.T. Mellon, Icarus **168**, 285–304 (2004)
A.M. Hessler, D.R. Lowe, R.L. Jones, D.K. Bird, Nature **428**, 736–8 (2004)
B.A. Hofmann, J.D. Farmer, Planet. Space Sci. **48**, 1077–1086 (2000)
B.A. Hofmann, Y. Krüger, A.E. Fallick, M. Eggimann, M.J. Van Kranendonk, Proc. 6th European Workshop on Astrobiology, Lyon, 16–18 October 2006, Ecole Normale Superieure, Lyon, France, 2006
H.J. Hofmann, K. Grey, A.H. Hickman, R.I. Thorpe, Geol. Soc. Am. Bull. **111**, 1256–1262 (1999)
N. Hoffman, P.R. Kyle, The ice towers of Mt. Erebus as analogues of biological refuges on Mars. Presented at 6th International Mars Conference, #3105 (2003)
N. Hoffman, Modern geothermal gradients on Mars and implications for subsurface liquids. Conference on the Geophysical Detection of Subsurface Water on Mars (2001)
N.G. Holm, M. Dumont, M. Ivarsson, C. Konn, Geochem. Trans. **7**, 7 (2004). doi:10.1186/1467-4866-7-7
A. Hynes, Earth Planet. Sci. Lett. **185**, 161–172 (2001)
B.A. Ivanov, D.D. Badukov, O.I. Yakovlev, M.V. Gerasimov, Yu.P. Dikov, K.O. Pope, A.C. Ocampo, in *The Cretaceous-Tertiary Event and other Catastrophes in Earth History*, ed. by G. Ryder, D. Fastovsky and S. Gartner. Special paper No 307 (Geological Society of America, 1996), pp. 125–139
R. Jaumann, Die Erosionsmorphologie des Mars: Genese, Verteilung und Stratigraphie von Erosionsformen und deren klimatische Bedeutung, DLR Forschungsbericht **2003–20**, pp. 261 (2003)
R. Jaumann, E. Hauber, J. Lanz, H. Hoffmann, G. Neukum, in *Astrobiology*, ed. by G. Horneck, C. Baumstark-Kahn (Springer, Berlin, 2002), pp. 89–109.
R. Jaumann, D. Reiss, S. Frei, G. Neukum, F. Scholten, K. Gwinner, T. Roatsch, K.-M. Matz., E. Hauber, U. Köhler, J.W. Head, H. Hiesinger, M. Carr, Geophys. Res. Lett. **32**, L16203 (2005). doi:10.1029/2002GL023415
A. Kappler, B. Schink, D.K. Newman, Geobiology **3**, 235–245 (2005)
J.S. Kargel, R.G. Storm, Anc. Glaciat. Mars Geol. **20**, 3–7 (1992)
K. Kashevi, D.R. Lovley, Science **301**, 934 (2003)
J.F. Kasting, Astrophys. Space Sci. **241**, 3–24 (1996)
J.F. Kasting, Origins of Life **27**, 291–307 (1997)
J.F. Kasting, D.P. Whitmire, R.T. Reynolds, Icarus **101**, 108–128 (1993)
P. Kharecha, J. Kasting, J. Siefert, Geobiology **3**, 53–76 (2005)

H.H. Kieffer, P. Zent, in *Mars*, ed. by H.H. Kieffer, B.M. Jakosky, C.W. Snyder, M.S. Matthews (University of Arizona Press, Tucson, 1992), pp. 1180–1220
G. Klingelhöfer, R.V. Morris, B. Bernhardt, C. Schröder, D.S. Rodionov, P.A. de Souza Jr., A. Yen, R. Gellert, E.N. Evlanov, B. Zubkov, J. Foh, U. Bonnes, E. Kankeleit, P. Gütlich, D.W. Ming, F. Renz, T. Wdowiak, S.W. Squyres, R.E. Arvidson, Science **306**, 1740–1746 (2004)
K.O. Konhauser, V.R. Phoenix, S.H. Bottrell, D.G. Adams, I.M. Head, Sedimentology **48**, 415–433 (2001)
M. Kretzschmar, Facies **7**, 237–260 (1982)
Y. Langevin, F. Poulet, J.-P. Bibring, B. Gondet, Science **307**, 1584–1586 (2005)
A. Lenardice, F. Nimmo, L. Moresi, J. Geophys. Res. **109**, E02003 (2004). doi:1029/2003JE002172
D.R. Lowe, G.R. Byerly, F.T. Kyte, A. Shukulyukov, F. Asaro, A. Krull, Astrobiology **3**, 7–48 (2003)
A.M. Macgregor, South Afr. J. Sci. **24**, 155–172 (1927)
M.C. Malin, K.S. Edgett, Science **290**, 1927–1937 (2000a)
M.C. Malin, K.S. Edgett, Science **288**, 2330–2335 (2000b)
M.C. Malin, K.S. Edgett, L.V. Posiolova, S.M. McColley, E.Z. Noe Dobrea, Science **314**, 1573–1577 (2006)
P. Masson, M.H. Carr, F. Costard, R. Greeley, E. Hauber, R. Jaumann, in *Chronology and Evolution on Mars* ed. by R. Kallenbach, J. Geiss, W.K. Hartmann. Space Science Series of ISSI. Space Sci. Rev. **96**, 333–364 (2001)
J.P. Mckinley, T.O. Stevens, F. Westall, Geomicrobiol. J. **17**, 43–54 (2000)
H.Y. McSween Jr., Meteoritics **29**, 757–779 (1994)
H.Y. McSween Jr., S.L. Murchie, J.A. Crisp, N.T. Bridges, R.C. Anderson, J.F. Bell III, D.T. Britt, J. Brückner, G. Dreibus, T. Economou, A. Ghosh, M.P. Golombek, J.P. Greenwood, R.J. Johnson, H.J. Moore, R.V. Morris, T.J. Parker, R. Rieder, R. Singer, H. Wänke, J. Geophys. Res. **104**(E4), 8679–8715 (1999)
H.Y. McSween, R.E. Arvidson, J.F. Bell III, D. Blaney, N.A. Cabrol, P.R. Christensen, B.C. Clark, J.A. Crisp, L.S. Crumpler, D.J. DesMarais, J.D. Farmer, R. Gellert, A. Ghosh, A. Gorevan, T. Graff, J. Grant, L.A. Haskin, K.E. Herkenhoff, J.R. Johnson, B.L. Jolliff, G. Klingelhoefer, A.T. Knudson, S. McLennan, K.A. Milam, J.E. Moersch, R.V. Morris, R. Rieder, S.W. Ruff, P.A. de Souza Jr., S.W. Squyres, H. Wänke, A. Wang, M.B. Wyatt, A. Yen, J. Zipfel, Science **305**, 842–845 (2004)
C.E. Mire, J.A. Tourjee, W.F. O'Brien, K.V. Ramanujachary, G.B. Hecht, Appl. Environ. Microbiol. **70**, 855–864 (2004)
D. Möhlmann, Icarus **168**, 318–323 (2004)
R.V. Morris, G. Klingelhöfer, B. Bernhardt, C. Schröder, D.S. Rodionov, P.A. de Souza Jr., A. Yen, R. Gellert, E.N. Evlanov, J. Foh, E. Kankeleit, P. Gütlich, D.W. Ming, J.F. Mustard, F. Poulet, A. Gendrin, J.-P. Bibring, Y. Langevin, B. Gondet, N. Mangold, G. Belucci, F. Altieri, Science **307**, 1594–1597 (2005)
D.P. Moser, T.M. Gihring, F.J. Brockman, J.K. Fredrickson, D.L. Balkwill, M.E. Dollhopf, B.S. Lollar, L.M. Pratt, E. Boice, G. Southam, G. Wanger, B.J. Baker, S.M. Pfiffner, L.-H. Lin, T.C. Onstott, Appl. Environ. Microbiol. **71**, 8773–8783 (2005)
L.M. Mukhin, M.V. Gerasimov, E.N. Safonova, Nature **340**, 46–48 (1989)
J.F. Mustard, F. Poulet, A. Gendrin, J.-P. Bibring, Y. Langevin, B. Gondet, N. Mangold, G. Belucci, F. Altieri, Science **307**, 1594–1597 (2005)
G. Neukum, R. Jaumann, H. Hoffmann, E. Hauber, J.W. Head, A.T. Basilevsky, B.A. Ivanov, S.C. Werner, S. van Gasselt, J.B. Murray, T. McCord, the HRSC Co-Investigator Team, Nature **432**, 971–979 (2004)
F. Nimmo, D. Stevenson, J. Geophys. Res. **105**, 11969–11979 (2000)
E.G. Nisbet, *The Young Earth* (G. Allen and Unwin, London, 1987) 402 pp.
E.G. Nisbet, M.J. Cheadle, N.T. Arndt, M.J. Bickle, Lithos **30**, 291–307 (1993)
E.G. Nisbet, N.H. Sleep, Nature **409**, 1083–1091 (2001)
B. Orberger, V. Rouchon, F. Westall, S.T. de Vries, C. Wagner, D.L. Pinti, in *Processes on the Early Earth*, vol. 405, ed. by W.U. Reimold, R. Gibson (Geol. Soc. Amer. Spec. Pub., 2006), pp. 133–152
R.J. Parkes, B.A. Cragg, S.J. Bale, J.M. Getliff, K. Goodman, P.A. Rochelle, J.C. Fry, A.J. Weightman, S.M. Harvey, Nature **371**, 410–413 (1994)
R.J. Parkes, G. Webster, B.A. Cragg, A.J. Weightman, C.J. Newberry, T.G. Ferdelman, J. Kallmeyer, B.B. Jørgensen, I.W. Aiello, J. Fry, Nature **436**, 390–394 (2005)
A.A. Pavlov, J.F. Kasting, L.L. Brown, K.A. Rages, R. Freedman, J. Geophys. Res. **105**, 11981–11990 (2001)
K. Pedersen, FEMS Microbiol. Lett. **185**, 9–16 (2000)
V.R. Phoenix, D.G. Adams, K.O. Konhauser, Chem. Geol. **169**, 329–338 (2000)
V.R. Phoenix, K.O. Konhauser, D.G. Adams, Geology **29**, 823–826 (2001)
E. Pierazzo, H.J. Melosh, Meteorit. Planet. Sci. **35**, 117–130 (2000)
D. Pieri, Science **210**, 895–897 (1980)
D.L. Pinti, K. Hashizume, Precambr. Res. **105**, 85–88 (2001)
D.L. Pinti, K. Hashizume, J. Matsuda, Geochim. Cosmochim. Acta **65**, 2301–2315 (2001)
R.G. Prinn, B. Fegley Jr., Earth Planet. Sci. Lett. **83**, 1–15 (1987)
N.E. Putzig, M.T. Mellon, K.A. Kretke, R.E. Arvidson, Icarus **173**, 325–341 (2005)

D. Reiss, R. Jaumann, Geophys. Res. Lett. **30**(6) (2003). doi:1029/2002GL016704
A. Ricardo, M.A. Carrigan, A.N. Olcott, S.A. Benner, Science **303**, 196 (2004)
R. Rieder, H. Wänke, T. Economou, A. Turkevich, J. Geophys. Res. **102**, 4027–4044 (1997a)
R. Rieder, T. Economou, H. Wänke, A. Turkevich, J. Crisp, J. Brückner, G. Dreibus, H.Y. McSween Jr., Science **278**, 1771–1774 (1997b)
R. Rieder, R. Gellert, R.C. Anderson, J. Brückner, B.C. Clark, G. Dreibus, T. Economou, G. Klingelhöfer, G.W. Lugmair, D.W. Ming, S.W. Squyres, C. d'Uston, H. Wänke, A. Yen, J. Zipfel, Science **306**, 1746–1749 (2004)
J.R. Rogers, P.C. Bennett, Chem. Geol. **203**, 91–108 (2004)
J. Sackmann, A.I. Boothroyd, K.E. Kraemer, Astrophys. J. **418**, 457–468 (1993)
M.A. Saito, D.M. Sigman, F.M.M. Morel, Inorganica Chimica Acta **356**, 308–318 (2003)
S. Schultze-Lam, G. Harauz, T.J. Beveridge, J. Bacteriol. **174**, 7971–7981 (1992)
D.H. Scott, J.M. Dohm, Martian highland channels: An age reassessment, Lunar Planet. Sci. Conf. XXIII, LPI, Houston, 1992, pp. 1251–1252
Y. Sekine, S. Sugita, T. Kadono, T. Matsui, J. Geophys. Res. **108**(E7), 5070 (2003). doi:10.1029/2002JE002034
R.P. Sharp, M.C. Malin, Geol. Soc. Am. Bull. **86**, 593–609 (1975)
C. Siebert, J.D. Kramers, T. Meisel, P. Morel, T.F. Nagler, Geochimica Cosmochimica Acta **69**, 1787–1801 (2005)
N.H. Sleep, J. Geophys. Res. **99**, 5639–5655 (1994)
N.H. Sleep, Annu. Rev. Earth Planet. Sci. **33**, 369–393 (2005)
N.H. Sleep, K.J. Zahnle, J. Geophys. Res. **103**, 28529–28544 (1998)
N.H. Sleep, K. Zahnle, P.S. Neuhoff, Proc. Nat. Acad. Sci. **98**, 3666–3672 (2001)
A.P. Sommer, D.S. McKay, N. Ciftcioglu, U. Oron, A.R. Mester, E.O. Kajander, J. Proteome Res. **2**, 441–443 (2003)
T. Spohn, M.H. Acuña, D. Breuer, M. Golombek, A. Greeley, A. Halliday, E. Hauber, R. Jaumann, F. Sohl, in *Chronology and Evolution on Mars*, ed. by R. Kallenbach, J. Geiss, W.K. Hartmann. Space Science Series of ISSI. Space Sci. Rev. **96**, 231–261 (2001)
S.W. Squyres, M.H. Carr, Science **231**, 249–252 (1986)
S.W. Squyres, J. Grotzinger, R.E. Arvidson, J.F. Bell III, W. Calvin, P.R. Christensen, B.C. Clark, J.A. Crisp, W.H. Farrand, K.E. Herkenhoff, J.R. Johnson, G. Klingelhöfer, A.H. Knoll, S.M. McLennan, H.Y. McSween Jr., R.V. Morris, J.W. Rise Jr., R. Rieder, L.A. Soderblom, Science **306**, 1709–1714 (2004)
L.J. Stahl, in *Biostabilization of sediments*, ed. by W.E. Krumbein (Biblioteks und Informationssystem der Carl von Ossietzky Universität, Oldenburg, 1994), pp. 41–54
T. Stevens, FEMS Microbiol. Rev. **20**, 327–337 (1997)
T.O. Stevens, J.P. Mckinley, Science **270**, 450–454 (1995)
T.O. Stevens, J.P. Mckinley, J.K. Fredrickson, Microb. Ecol. **25**, 35–50 (1993)
D.J. Stevenson, Nature **400**, 32 (1999)
B.M. Tebo, J.R. Bargar, B.G. Clement, G.J. Dick, K.J. Murray, D. Parker, R. Verity, S.M. Webb, Annu. Rev. Earth Planet. Sci. **32**, 287–328 (2004)
M.M. Tice, D.R. Lowe, Nature **431**, 549–552 (2004)
J.K.W. Toporski, A. Steele, F. Westall, K.L. Thomas-Keprta, D.S. McKay, Astrobiology **2**, 1–26 (2002)
N.H. Trewin, A.H. Knoll, Palaios **14**, 288–294 (1999)
Y. Ueno, H. Yoshioka, S. Maruyama, Y. Isozaki, Geochim. Cosmochim. Acta **68**, 573–589 (2004)
J.W. Valley, W.H. Peck, E.M. King, S.A. Wilde, Geology **30**, 351–354 (2002)
P. Van Thienen, N.J. Vlaar, A.P. van der Berg, Phys. Earth Planet. Int. **142**, 61–74 (2004)
M. van Zuilen, A. Lepland, G. Arrhenius, Nature **418**, 627–630 (2002)
M.A. van Zuilen, K. Mathew, B. Wopenka, A. Lepland, K. Marti, A. Arrhenius, Geochim. Cosmochim. Acta **69**, 1241–1252 (2005)
J.C.G. Walker, Nature **329**, 710–712 (1987)
M.M. Walsh, Precambr. Res. **54**, 271–293 (1992)
M.M. Walsh, Astrobiology **4**, 429–437 (2004)
P.H. Warren, J. Geophys. Res. **98**, 5335–5345 (1993)
A.J. Watson, T.M. Donahue, W.R. Kuhn, Earth Planet. Sci. Lett. **68**, 1–6 (1984)
E.B. Watson, T.M. Harrison, Science **308**, 841–844 (2005)
B.P. Weiss, Y.L. Yung, K.H. Nealson, Proc. Natl. Acad. Sci. **97**, 1395–1399 (2000)
S.C. Werner, Major aspects of the chronostratigraphy and geologic evolutionary history of Mars. PhD Thesis, Free University Berlin, 2005
F. Westall, Palevol **2**, 485–501 (2003)
F. Westall, G. Southam, in *Archean Geodynamics and Environments*, ed. by K. Benn et al. AGU Geophys. Monogr., vol. 164 (2006), pp. 283–304

F. Westall, L. Boni, E. Guerzoni, Paleontology **38**, 495–528 (1995)
F. Westall, S.T. de Vries, W. Nijman, V. Rouchon, B. Orberger, V. Pearson, J. Watson, A. Verchovsky, I. Wright, J.-N. Rouzaud, D. Marchesini, S. Anne, in *Processes on the Early Earth*, ed. by W.U. Reimold, R. Gibson, Geol. Soc. Amer. Spec. Pub., vol. 405 (2006a), pp. 105–131
F. Westall, C.E.J. de Ronde, G. Southam, N. Grassineau, M. Colas, C. Cockell, H. Lammer, Phil. Trans. Roy. Soc. B **185**, 1857–1875 (2006b)
F. Westall, L. Lemelle, M. Salomé, A. Simionovici, G. Southam, C.E.J. Ronde, N. Grassineau, M. Colas, B. Orberger, Geochemical profiling of an Early Archean, littoral environment, photosynthetic microbial mat from the Barberton greenstone belt, South Africa. EANA, Lyon 16–18 October 2006c, Abstr.
J. Wierzchos, L.G. Sancho, C. Ascaso, Environ. Microbiol. **7**, 566–575 (2005)
S.A. Wilde, J.W. Valley, W.H. Peck, C.M. Graham, Nature **409**, 175–178 (2001)
O.I. Yakovlev, Yu.P. Dikov, M.V. Gerasimov, Geochem. Int. **30**(7), 1–10 (1993)
O.I. Yakovlev, Yu.P. Dikov, M.V. Gerasimov, F. Wlotzka, J. Huth, in *Lunar and Planetary Science XXXIII*, Abstract #1271 (Lunar and Planet. Inst., Houston, 2002) (CD-ROM)
A.S. Yen, R. Gellert, C. Schröder, R.V. Morris, J.F. Bell III, A.T. Knudson, B.C. Clark, D.W. Ming, J.A. Crisp, R.E. Arvidson, D. Blaney, J. Brückner, P.R. Christensen, D.J. DesMarais, P.A. de Souza Jr., T.E. Economou, A. Ghosh, B.C. Hahn, K.E. Herkenhoff, L.A. Haskin, J.A. Hurowitz, B.L. Jolliff, J.R. Johnson, G. Klingelhöfer, M.B. Madsen, S.M. McLennan, H.Y. McSween, L. Richter, R. Rieder, D. Rodionov, L. Soderblom, S.W. Squyres, N.J. Tosca, A. Wang, M. Wyatt, J. Zipfel, Nature **436**, 49–54 (2005)
K.J. Zahnle, N.H. Sleep, in *Comets and the Origin and Evolution of Life*, 1st edn., ed. by P.J. Thomas, C.F. Chyba, C.P. McKay (Springer, New York, 1997), pp. 175–208
K.J. Zahnle, N.H. Sleep, in *Comets and the Origin and Evolution of Life*, 2nd edn., ed. by P.J. Thomas, R.D. Hicks, C.F. Chyba, C.P. McKay (Springer, Berlin, 2006), pp. 207–252
M.Y. Zolotov, E.L. Shock, Meteorit. Planet. Sci. **35**, 629–638 (2000)