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REVIEW ARTICLE

What makes a planet habitable?

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Abstract This work reviews factors which are important for the evolution of habitable Earth-like planets such as the effects of the host star dependent radiation and particle fluxes on the evolution of atmospheres and initial water inventories. We discuss the geodynamical and geophysical environments which are necessary for planets where plate tectonics remain active over geological time scales and for planets

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which evolve to one-plate planets. The discoveries of methane–ethane surface lakes on Saturn's large moon Titan, subsurface water oceans or reservoirs inside the moons of Solar System gas giants such as Europa, Ganymede, Titan and Enceladus and more than 335 exoplanets, indicate that the classical definition of the habitable zone concept neglects more exotic habitats and may fail to be adequate for stars which are different from our Sun. A classification of four habitat types is proposed. Class I habitats represent bodies on which stellar and geophysical conditions allow Earth-analog planets to evolve so that complex multi-cellular life forms may originate. Class II habitats includes bodies on which life may evolve but due to stellar and geophysical conditions that are different from the class I habitats, the planets rather evolve toward Venus- or Mars-type worlds where complex life-forms may not develop. Class III habitats are planetary bodies where subsurface water oceans exist which interact directly

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with a silicate-rich core, while class IV habitats have liquid water layers between two ice layers, or liquids above ice. Furthermore, we discuss from the present viewpoint how life may have originated on early Earth, the possibilities that life may evolve on such Earth-like bodies and how future space missions may discover manifestations of extraterrestrial life.

Keywords Habitability · Origin of life · Terrestrial planets · Subsurface oceans · Atmosphere evolution · Earth-like exoplanets · Space weather · Astrobiology

1 Introduction: requirements for habitability

Discussing the habitability of a celestial body is very closely connected to the discussion of life outside Earth. Life of course cannot ignore its connection to habitability, but the reverse is not automatically true. The word "habitable" means "suitable to live in (or on)", and life cannot originate without habitability. But as an abandoned house might be perfectly suitable to live in (habitable) this does not necessarily mean there is somebody living in it. The question to ask in regard to celestial bodies is not only "Is it habitable?", but also "Could life have originated and evolved there?"

Besides the potential habitats in the Solar System more than 335 giant exoplanets have been detected[1] to date. The current status of exoplanet characterization shows a surprisingly diverse set of mainly giant planets. Some of their properties have been measured using photons from the host star, a background star, or a mixture of the star and planet. These indirect techniques include radial velocity, micro-lensing, transits, and astrometry (Beaugé et al. 2008; Fridlund and Kaltenegger 2008; Rauer and Erikson 2008). Earth is until now the only example of a known habitable planet. Compared to other terrestrial planets in our Solar System, Earth is unique: it has liquid water on its surface, an atmosphere with a greenhouse effect that keeps its surface above freezing, and the right mass to maintain tectonics (e.g., Kasting et al. 1993). Earth orbits its host star—our Sun—within a region that is called the habitable zone (HZ)—the region where an Earth analog planet can maintain liquid water on its surface is shown in Fig. 1.

The classical concept of the HZ was first proposed by Huang (1959, 1960) and has been modeled by several authors (e.g., Rasool and deBergh 1970; Hart 1979; Kasting et al. 1993). The differences in the calculations are the climatic constraints imposed on the limits of the HZ. In all cases the stellar habitable zone is a spherical shell around a main sequence star where a planet with an atmosphere can support liquid water at a given time. The width and distance of this shell depends on the stellar luminosity that evolves during the star's lifetime. The continuously habitable zone (CHZ) has been introduced as the zone that remains habitable around a star during a given period of time (Hart 1978).

Liquid water seems to be an important requirement for habitability. Liquid water has been pointed out as the best solvent for life to emerge and evolve in. Some of the important characteristics of liquid water as a solvent include: a large dipole moment,

[1] http://www.exoplanet.eu.

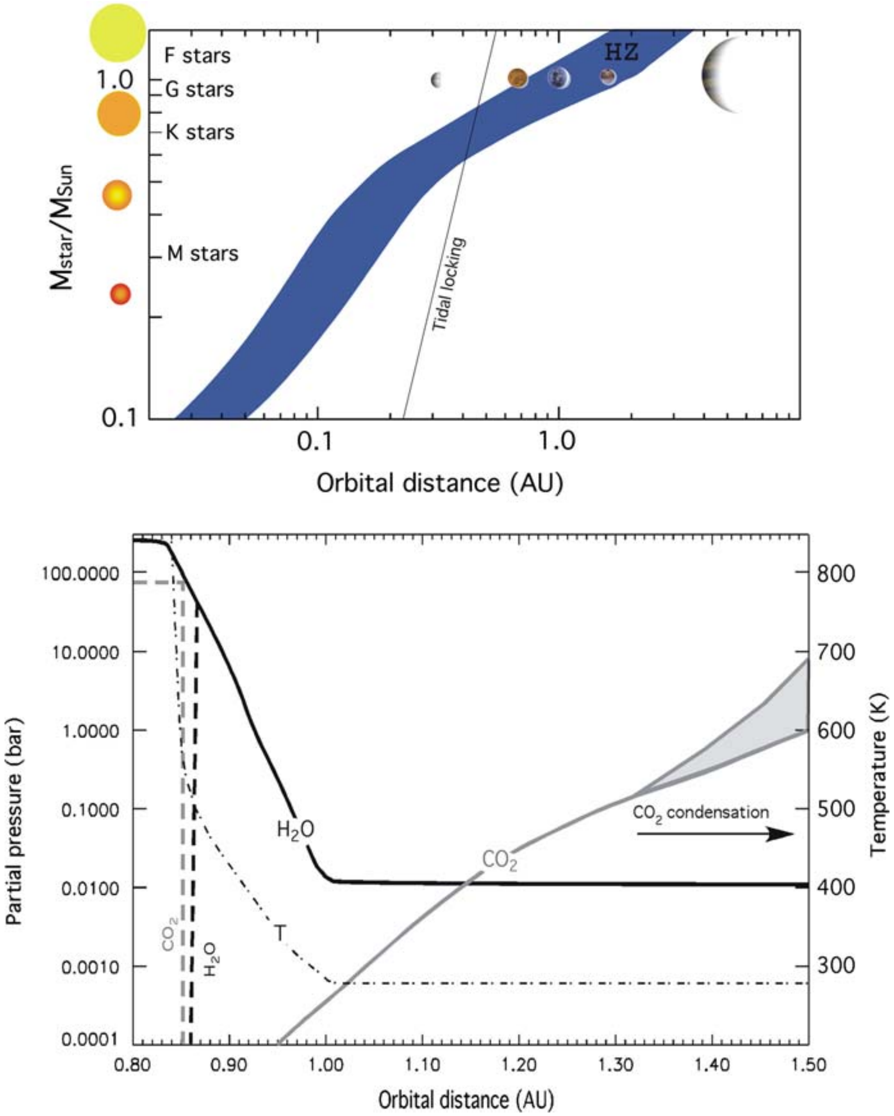


Fig. 1 The HZ (*upper panel*) and the chemistry composition (*lower panel*) of an Earth-analog atmosphere as a function of distance from its host star. The *dashed-dotted line* represents the surface temperature of the planet and the *dashed lines* correspond to the inner edge of the HZ where the greenhouse conditions vaporize the whole water reservoir (adapted from Kaltenegger and Selsis 2007)

the capability to form hydrogen bonds, to stabilize macromolecules, to orient hydrophobic–hydrophilic molecules, etc. Water is an abundant compound in our galaxy, it can be found in different environments from cold dense molecular clouds to hot stellar atmospheres (e.g., Cernicharo and Crovisier 2005). Water is liquid at a large range of temperatures and pressures and it is a strong polar–nonpolar solvent. This dichotomy

Table 1 Limits for life as we know it (Cockell 1999)

Parameters	Limits for life as we know it
Temperature (°C)	<0–113
Surface pressure (MPa)	>100
Acidity (pH)	>12
UV radiation	On early Earth 200–400 nm range of the 25% less luminous Sun reached the surface
Atmospheric composition	Pure CO_2 can be tolerated by some organisms. High N_2 will not prevent life
Water availability, liquid H_2O on the surface	Liquid H_2O should be present but some halophilic organisms live in high (4–5 M) NaCl

is essential for maintaining stable biomolecular and cellular structures (DesMarais et al. 2002) and there are a large number of organisms that are capable of living in water.

The inner Edge of the HZ is the distance where runaway greenhouse conditions vaporize the whole water reservoir and, as a second effect, induce the photodissociation of water vapor and the loss of hydrogen to space, see (McKay and Stoker 1989; Kulikov et al. 2007; Lammer et al. 2008; Tian et al. 2008) for a detailed discussion. The outer edge of the HZ is the distance from the star where a maximum greenhouse effect fails to keep the surface of the planet above the freezing point, or the distance from the star where CO_2 starts condensing (Kasting et al. 1993).

On an Earth-like planet where the carbonate–silicate cycle is at work, the level of CO_2 in the atmosphere depends on the orbital distance: CO_2 is a trace gas close to the inner edge of the HZ but a major compound in the outer part of the HZ. Earth-like planets close to the inner edge are expected to have a water-rich atmosphere or to have lost their water reservoir to space.

The classical HZ is defined for surface conditions only, chemolithotrophic life, a metabolism that does not depend on the stellar light, can still exist outside the HZ, thriving in the interior of the planet where liquid water is available. Such metabolisms (the ones known on Earth) do not produce O_2 and relay on very limited sources of energy (compared to stellar light). They mainly catalyze reactions that would occur at a slower rate in purely abiotic conditions and they are thus not expected to modify a whole planetary environment in a detectable way.

From our experience on Earth we know that if life once originated, it seems that it can adapt to all kinds of extreme environmental conditions (e.g., Rothschild and Mancinelli 2001) as shown in Table 1.

The basic habitability requirements for life as we know are constrained by the following main factors:

– a certain time span where a celestial body can accumulate enough building blocks necessary for the origin of life,
– liquid water which is in contact with these building blocks,

– external and internal environmental conditions that allow liquid water to exist, on a celestial body over a time span necessary that allows life to evolve.

After life has originated, depending on the evolution of the planetary environment, it may evolve to form complex multi-cellular life forms or it may remain as microbial life which may adapt to extreme environmental planetary conditions.

Under such considerations planets with eccentric orbits near or within the HZ may also be potential habitable worlds (Williams and Pollard 2002) due to climate stability in the long-term because the crucial parameter is the average stellar flux received by the planet during an entire orbit.

Other possible exotic habitable environments, like extrasolar giant planets' satellites (e.g., Europa- or Titan-like environments) located outside the classical HZ, have not been discussed as targets in the search for life with a terrestrial planet finding mission, because its related biomarkers will not be detectable remotely. Life that does not influence the atmosphere of its planet on a global scale (like on Earth) will not be remotely detectable. Therefore, such potential habitats can only be investigated in the Solar System with appropriate space missions.

The HZ concept is still evolving (e.g., Segura and Kaltenegger 2008) as we learn more about planetary formation, evolution and improve radiative transfer and 3-D atmospheric models that allow more accurate calculations of a planet's temperature profile, as well as the relation and interaction between the host star radiation and plasma environment and the planet.

The question of what makes a planet habitable is much more complex than having a planet located at the right distance from its host star so that water can be liquid on its surface. Furthermore, various geophysical and geodynamical aspects, the radiation and the host stars plasma environment can influence the evolution of Earth-type planets and life if it originated (Scalo et al. 2007; Khodachenko et al. 2007b; Lammer et al. 2007). For instance, a subsurface ocean within the satellite of a gas giant may be habitable for some life form, albeit not necessarily Earth-analog life. On the other hand there may be many habitable exoplanets, but life could never evolve from microorganisms to complex multi-cellular life. With the discovery of planets beyond the Solar System and the search for life in exotic environments such as Mars, Europa, Titan and Enceladus, the various habitats need a separate definition.

In this work we try to focus on aspects which are relevant to habitability but were so far not addressed in detail in studies related to HZs. In Sect. 2 we discuss the sources of life and its building blocks, which are ingredients of the planetary nebulae and the planetary environments which form out of them. As shown in Fig. 1, planets within the classical HZ orbit at different distances around different stellar spectral-type stars. Because planets evolve together with their host stars, we discuss the difference and expected consequences of the radiation and particle environment of F, G, K and M stars in Sect. 3. In Sect. 4 we focus on factors which are relevant for the evolution of habitable terrestrial planets, but are not discussed in detail in the classical HZ concept. These factors contain basic geophysical conditions in relation to their host stars, the role of plate tectonics, magnetic dynamos, ionosphere boundaries, induced magnetic fields, as well as the protection of planetary atmospheres against stellar plasma flows.

In Sect. 5 we define four habitat classes. In class I the stellar and geophysical conditions of Earth-like (analog) planets where surface life as we know it can evolve are represented. Class II habitats will focus on planets where in the beginning life may evolve because these planets start an evolution similar to class I types but due to different stellar and geophysical conditions the planetary environments and life evolve different than on Earth. Class III habitats are planetary bodies with subsurface water layers in the form of subsurface oceans which interact with silicates (e.g., Europa). Finally, class IV habitats have liquid water layers between two ice layers or liquids above ice (e.g., Ganymede, Callisto, Enceladus, Titan-lakes, and "Ocean planets"). After discussing and defining these habitability classifications, in Sect. 6 we give an overview of the present knowledge how life may have originated on early Earth, and the possible characteristics of habitable planetary bodies which fall in the four habitat categories. Finally, in Sect. 7 we discuss future space missions which may be helpful to investigate the habitats addressed in this work.

2 Cosmo-chemical aspects for the building blocks of life

Astronomical observations in the last decades have identified gaseous molecules and solids including carbon chains, aromatic hydrocarbons, carbonaceous grains and carbon-bearing ices that are observed in different galactic and extragalactic regions (Henning and Salama 1998). It is interesting to note that the carbon chemistry in different space environments proceeds to form many carbonaceous molecules—small and probably very large—which are common on Earth (Ehrenfreund and Spaans 2007). Where are those compounds formed and how do they evolve?

2.1 Interstellar clouds

The interstellar medium (ISM) hosts an active carbon chemistry that is strongly influenced by environmental conditions such as density, temperature and radiation conditions (Wooden et al. 2004; Snow and McCall 2006). Table 2 lists the parameters for interstellar clouds. Molecular production is most effective in diffuse and dense interstellar clouds.

The ISM comprises a few % of the mass of the galaxy. Its main constituent is H and He gas. However, submicron dust particles, mainly composed of silicates and carbonaceous material, are also present in small concentrations ($\sim$1% relative to interstellar gas). Dust can adsorb ice mantles in cold interstellar regions. In the so-called dark interstellar clouds, characterized by high densities (see Table 2) active chemical pathways lead to complex carbon molecules in the gas phase and solid state (Ehrenfreund and Charnley 2000).

Dark clouds offer a protected environment for the formation of larger molecules. Those regions experience a limited radiation field of $\sim 10^3$ photons cm^{-2} s^{-1} induced by cosmic rays (Prasad and Tarafdar 1983). Icy surfaces of small dust grains offer a surface for catalytic reactions to form carbon bearing ice molecules such as CO, CO_2,

Table 2 Phases of the interstellar medium (adapted from Wooden et al. 2004)

ISM component	Designation	Temperature (K)	Density (cm^{-3})
Hot ionized medium	Coronal gas	10^6	0.003
Warm ionized medium	Diffuse ionized gas	10^4	>10
Warm neutral medium	Intercloud HI	10^4	0.1
Atomic cold neutral medium	Diffuse clouds	100	10–100
Molecular cold neutral medium	Dark clouds molecular clouds dense clouds	<50	10^3–10^5
Molecular hot cores	Protostellar cores	100–300	$>10^6$

CH_3OH and others (Gibb et al. 2004). Observations at infrared, radio, millimeter, and sub-millimeter frequencies show that a large variety of organic molecules are present in the dense interstellar gas (http://www.astrochemistry.net lists more than 150 molecules).

Among them are simple molecules, such as CO, the most abundant carbon-containing species, with a ratio of $CO/H_2 \sim 10^{-4}$, but nitriles, aldehydes, alcohols and hydrocarbon chains have also been identified. Furthermore, Kuan et al. (2003) recently claimed that they detected the amino acid glycine in “hot-cores” that represent high-density regions close to forming stars.

The diffuse ISM is characterized by lower density and higher temperatures (more than 100 K). Ices are not present in those regions and a strong radiation field of $\sim 10^8$ photons cm^{-2} s^{-1} (Mathis et al. 1983) dominates the formation and evolution of carbon compounds. Small carbonaceous molecules in the gas phase are easily destroyed by radiation. The identification of many small carbon chains in dense clouds implies that their destruction is well balanced by active formation routes.

Circumstellar envelopes are regarded as the largest factories of carbon chemistry in space where molecular synthesis occurs on short timescales (several hundred years, Kwok 2004). Acetylene, a molecule formed in abundance in such regions, seems to be a key species that drives the chemistry towards aromatic rings, stable carbon molecules which are especially resistant to radiation.

Cosmic abundances in the ISM are derived by measuring elemental abundances in stellar photospheres, the atmospheric layer just above the stellar surface. These cosmic elemental abundances determine the amount of elements available for the formation of molecules and particles. In dense interstellar clouds gaseous CO can account for $\sim$20% of the cosmic carbon. Figure 2 shows examples of carbonaceous matter present or anticipated in interstellar clouds. Several allotropes of carbon, among them diamonds, graphite and fullerenes have been identified in space environments, in particular in meteorites (Ehrenfreund and Charnley 2000; Pendleton and Allamandola 2002). Diamonds were proposed to be the carriers of the 3.4 and 3.5 µm emission bands observed in planetary nebulae. However, graphite is only present in meteorites and has not been identified in the ISM up to now.

The polyhedral C_{60} fullerene geometry proposed by Kroto et al. (1985) is related to the presence of fullerene compounds in interstellar space. Until now fullerenes have

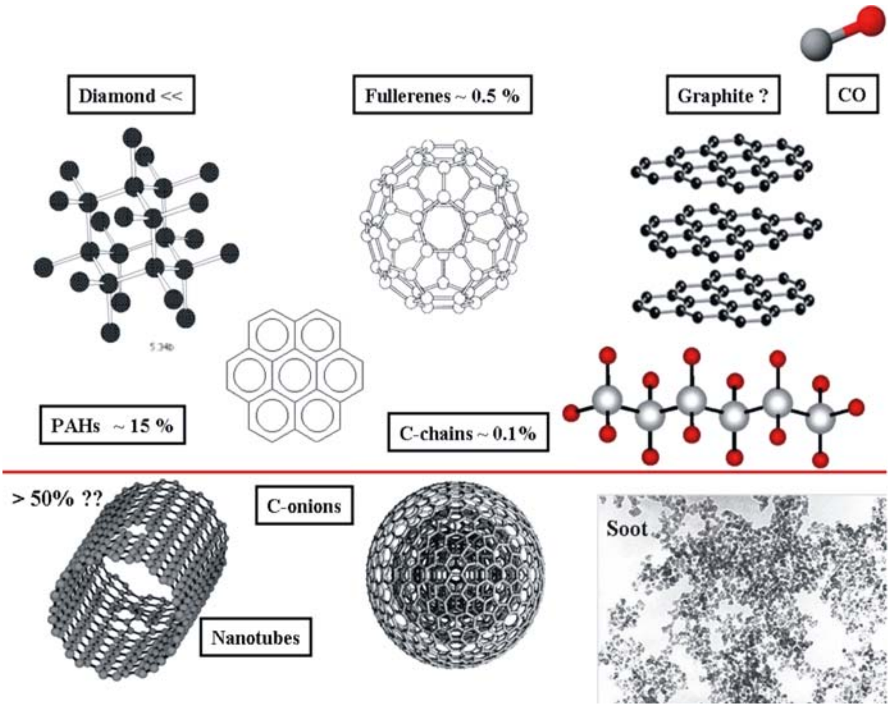


Fig. 2 Examples of carbonaceous matter present or anticipated in interstellar clouds. Several allotropes of carbon are found in space environment, e.g., in carbonaceous meteorites (diamonds, graphite, fullerenes, amorphous carbon). However, the presence of fullerenes and graphite in the interstellar medium is uncertain and the abundance of nano-diamonds is rather low. Aromatic hydrocarbons are the most abundant form of carbon in the universe. They exist as polycyclic aromatic hydrocarbons (PAHs) in the gas phase and are likely to be present in high abundance as aromatic networks (amorphous carbon) in carbonaceous grains

only been identified in meteorites (Becker and Bunch 1997). Possible fingerprints of the C_{60}^{+} ion were discovered in the near-infrared spectra of diffuse interstellar clouds (Foing and Ehrenfreund 1994; Ehrenfreund et al. 2006). Cosmic dust models indicate that the majority of carbon in diffuse clouds (up to 80%) is incorporated into carbonaceous grains and gaseous polycyclic aromatic hydrocarbon (PAHs). PAHs are ubiquitous in galactic and extragalactic regions and are the most abundant carbonaceous gas phase molecules in space (Allamandola et al. 1999; Smith et al. 2007). There is spectroscopic evidence that carbonaceous grains are predominantly made of amorphous carbon (Mennella et al. 1998; Henning et al. 2004). Other forms such as C-onions and nanotubes have been proposed as well. It is a crucial goal for astrochemistry to identify the nature of this abundant aromatic material.

2.2 From interstellar clouds to forming planetary systems

Interstellar clouds provide the raw material, namely gas and dust, for the formation of stars and planetary systems (Boss 2004). Astronomical observations of Solar System objects in combination with the analysis of extraterrestrial material (meteorites and interplanetary dust particles IDPs) provided us with important insights into the processes that occurred during planetary formation in our solar nebula. Our Solar System was formed about 4.6 Gyr ago through the gravitational collapse of an interstellar cloud. Infalling interstellar gas and dust were thoroughly mixed and processed in order to form Solar System bodies.

The small interstellar dust particles grew bigger and accreted more material, eventually forming planets or moons. The remaining material, not incorporated into planets, formed the populations of comets and asteroids that experienced a diverse history. A substantial fraction of small Solar System bodies impacted the young planets in the early history of the Solar System. The large quantities of extraterrestrial material delivered to young planetary surfaces during the heavy bombardment phase may have played a key role in life's origin (Chyba and Sagan 1992).

Recent data from the Stardust mission confirmed large-scale mixing in the solar nebula (Brownlee et al. 2006). The carbonaceous inventory of our Solar System is a mixture of materials including:

– highly processed material that was exposed to high temperature and radiation;
– newly formed compounds;
– relatively pristine material with strong interstellar heritage.

Future observations of organic compounds in our Solar System and the analysis of extraterrestrial material will provide us with more knowledge on the evolution of carbonaceous compounds and processes that occurred during the formation of our Solar System (Roush and Cruikshank 2004; Ehrenfreund et al. 2002).

In the inner Solar System where terrestrial planets formed no original carbon material (volatile or refractory) could have survived the high temperatures. Organic matter on terrestrial planets must have formed after the planetary surfaces had cooled down. The extraterrestrial delivery of organic material during impacts of small Solar System bodies may have delivered important carbon compounds (Ehrenfreund et al. 2002). In order to reconstruct scenarios for the origin of life a prerequisite is to quantify the intrinsic inventory of carbonaceous material on Earth as well as the extraterrestrial influx from comets and meteorites. Another crucial question will be to investigate how this organic material could have assembled to form complex bio-molecules.

Our universe provided carbon and dust in large amounts as early as 500 million years after the big bang (Spaans 2004). This indicates that the formation of planetary systems with rocky terrestrial planets may have occurred long before the formation of our own Solar System. Astronomical observations using better instrumentation allow us to reveal more details on the composition of interstellar dust and gas, comets and asteroids. These data and results from more sensitive techniques applied to the analysis of meteorites can determine more accurately the molecular inventory of material that was transported to young planets by exogenous delivery.

From the latest stage of cosmochemical discoveries we can summarize that the building blocks of life should be available on every planetary system. After the planets have formed the molecules necessary for life are delivered by meteorites and comets to these early planets. Depending on the environmental conditions and the time-span where these conditions (e.g., water, surface temperature, surface atmosphere pressure, etc.) remain non-destructive to the delivered molecules, life may origin like it did on early Earth.

3 Stellar activity, radiation and plasma environment of main sequence stars

One of the major questions of current and future exoplanet finding missions (CoRoT, Kepler, SIM, Darwin/TPF-C/-I, etc.) is which of the main sequence star-types (M, K, G, F) may be good or at least preferred candidates for hosting habitable terrestrial planets? Obviously, the search for Earth-like exoplanets should not be limited only to the Sun-like G-type stars (e.g., Scalo et al. 2007; Kaltenegger et al. 2009). It needs to be extended also to lower mass M- and K-type stars, as well as to slightly more massive F-type stars. These stars have masses lower than 2.0 M_{Sun}, where M_{Sun} is the solar mass, and lifetimes longer than 1 Gyr. 95% of the stars in the mass range between 0.1 and 2.0 M_{Sun}, are M-type dwarf stars ($M < 0.6M_{\mathrm{Sun}}$), which are also the most numerous stars in the galaxy. This makes the study of the main sequence low mass stars and their potential influence on the planetary environments very important to the characterization of exoplanets and their possible habitability (Lammer 2007; Segura et al. 2003).

The stellar X-ray and EUV (XUV) radiation and the stellar wind constitute a permanent forcing of the upper atmosphere of the planets, thereby affecting the atmospheric evolution and chances for life to emerge there (Kulikov et al. 2007; Lammer et al. 2008; Tian et al. 2008). The effect of these forcing terms is to ionize, heat, chemically modify, and slowly erode the upper atmosphere throughout the lifetime of a planet as shown in Fig. 3. The closer to the star the planet is, the more efficient are these processes. Protection of the upper atmosphere of a planet against stellar XUV/EUV and stellar wind factors requires a strong intrinsic dipole magnetic field (Khodachenko et al. 2007a,b). In general, the boundaries of the HZ shown in Fig. 1 change throughout the star's lifetime as the stellar luminosity and activity evolve. This evolution is different for different star types, and furthermore depends on their age and the location

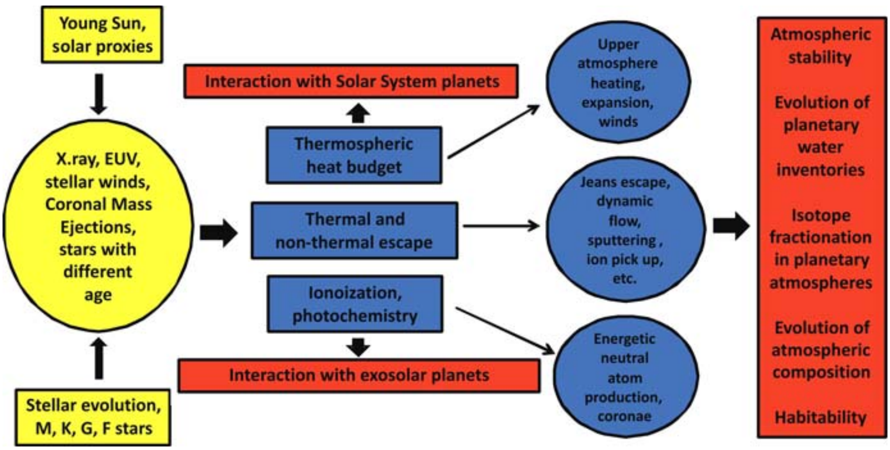


Fig. 3 Illustration on the radiation and plasma environment interactions with the upper atmosphere

of the HZ around the host star. This makes the type and age of a host star an important factor, which affects the evolution of potentially habitable worlds.

3.1 Radiation environment of lower mass stars

Activity in late-type stars (i.e., spectral types G, K, M) has been the subject of intense studies for many years. The relevant physical phenomena of stellar activity and their observational manifestations include modulations of the stellar photospheric light due to stellar spots, intermittent and energetic flares, coronal mass ejections (CMEs), stellar cosmic rays, enhanced coronal X-rays, and enhanced chromospheric UV emission (see Ayres 1997; Gershberg 2005; Scalo et al. 2007 and references therein).

Numerous observations of stars in clusters have revealed that single late-type stars spin down monotonically with their age which seems to roughly follow a power-law expression: $P_{\rm rot} \propto t^{-1/2}$ (e.g., Skumanich 1972; Newkirk 1980; Soderblom 1982; Ayres 1997) because of angular momentum loss. Earlier studies already pointed out a strong correlation between the rotation rate of a star and its activity level (Wilson 1966; Kraft 1967). This correlation shows interdependence between stellar activity and age (e.g., Keppens et al. 1995; Ribas et al. 2005). For solar-type stars this has been studied within the *Sun in Time* project (Ribas et al. 2005 and references therein). Based on a large number of multi-wavelength (X-ray, EUV, FUV, UV) observations of a homogeneous sample of single nearby G0–5 main sequence stars with known rotation periods and well-determined physical properties, luminosities and ages (covering 100 Myr–8.5 Gyr), it has been concluded that the solar integrated XUV flux (0.1–120 nm) was higher by a factor of six 3.5 Gyr ago than compared to the present state. Also, during the first 100 Myr after the Sun arrived at the zero-age main-sequence (ZAMS), the integrated XUV flux was up to 100 times more intense than today (Ribas et al. 2005). After this very active stage, the solar XUV flux quickly decreases with time following a power law relationship (t[Gyr]$^{-1.72}$). In addition to that, the results of Wood et al. (2002, 2004, 2005) and Holzwarth and Jardine (2007) indicate that young solar-type stars may have massive stellar winds connected to their mass loss rates, 100–1,000 times that of the present solar value.

Audard et al. (2000) found that the energy of flares correlates with the stellar activity (as characterized by the X-ray emission). The occurrence of energetic flares increases with increasing $\log(L_{\rm x}/L_{\rm bol})$ and then stays constant for saturated stars. Here, $L_{\rm x}$ is X-ray luminosity of a star, which is used as a proxy of its overall activity, and $L_{\rm bol}$ is bolometric luminosity. The time span of activity saturation is given in Table 3, which presents the evolution of $\log(L_{\rm x}/L_{\rm bol})$ with time for the stars of various masses, including solar-type and M-type stars (Scalo et al. 2007).

This activity–age portrait also shows that lower mass stars spend more time in the saturated plateau before beginning their activity decay. In particular, solar-type stars stay at saturated emission levels until they age to $\sim$100 Myr, and then their X-ray luminosities rapidly decrease as a function of age following a power law relationship $\propto (t[\mathrm{Gyr}])^{-1.72}$ (Ribas et al. 2005). M-type stars, on the other hand, have saturated emission periods up to 0.5–1 Gyr (and possibly longer for late M stars). After that the luminosity decreases in a way similar to solar-type stars.

Table 3 Time span in Gyr where $L_x/L_{bol(Sun)}$ as a function of stars with masses $\leq 1M_{Sun}$ where the $L_x/L_{bol(Sun)}$ is about 1,700 and ≥ 100 times larger than at the present Sun (after Scalo et al. 2007)

M_{Sun}	t [Gyr] for 1,700 $L_x/L_{bol(Sun)}$	t [Gyr] for $\geq 100L_x/L_{bol(Sun)}$
1.0	~0.05	~0.3
0.9	~0.1	~0.48
0.8	~0.15	~0.65
0.7	~0.2	~1.0
0.6	~0.3	~1.47
0.5	~0.5	~2.0
0.4	~0.75	~3.0
0.3	~1.0	~4.15
0.2	~1.58	~6.5
0.1	~4.6	>10.0

As a rule of thumb, the time evolution of M-type star X-ray luminosities could be considered as a constant luminosity $L_x \sim 7 \times 10^{28}$ erg s^{-1} up to the end of the saturation phase followed then by the power law decrease.

K type stars remain at active emission levels somewhat longer than G stars and afterwards the activity decreases by following a power law relationship. Early M-type stars appear to stay at high activity levels until ages of about 1 Gyr, and then decrease in an analogous way to G and K-type stars (Scalo et al. 2007; and references therein). In general, early K-type stars and early M stars may have XUV emissions of about 3–4 times and about 10–100 times higher, respectively, than solar-type G stars of the same age (Ribas et al. 2005).

In addition to that, partial or total tidal-locking of the terrestrial planets inside HZs of the low mass stars has long lasting consequences regarding plate-tectonics, generation of internal magnetic moments, and as a result the protection of the planetary atmosphere against erosion by stellar wind and CMEs. Weakening of the atmosphere-protective magnetosphere also results in easier penetration of galactic (Grießmeier et al. 2004, 2005, 2008) and stellar cosmic rays into the planetary atmosphere. This can lead to the kinds of ozone-depletion chemistry (Grenfell et al. 2007, 2008), similar to that on Earth in the polar regions, where energetic particles can more easily enter the upper atmosphere. If M star activity is accompanied by energetic particle emission, i.e., stellar cosmic rays, then the energetic particle flux due to the parent star could be orders of magnitude larger than that on the Earth. Similar effects can be expected for a tidally locked planet with a weak magnetic moment and therefore a stronger galactic cosmic ray exposure (Grießmeier et al. 2005, 2008).

3.2 Particle and plasma environment and habitability aspects

3.2.1 Stellar winds and Coronal Mass ejections

Along with the high and long lasting stellar XUV emissions and energetic particle fluxes, another crucial factor of stellar activity which may strongly affect potential habitability of a planet is the stellar wind plasma flow. The stellar plasma flow interacts with planetary magnetospheres, and in the case of weak planetary intrinsic magnetic fields, the stellar plasma flow can erode the planetary atmosphere and surface. Important parameters for characterization of the "stellar plasma–planetary atmosphere/surface interaction" are the stellar wind density and velocity, which are highly variable with stellar age and depend on the stellar spectral type as well as on the orbital distance of the planet. We currently have a good knowledge of these characteristics in the Sun, whereas the amount of data of this kind related to other stars is much more limited.

While the plasma properties of the solar wind can be measured directly with satellites, the situation is more difficult for other stars. The dense stellar winds from hot stars, cool giants/supergiants and young T-Tauri stars produce spectroscopic features from which the wind parameters can be derived. The tenuous winds from solar-like stars, on the other hand, are a few orders of magnitude less massive and thus cannot be detected spectroscopically.

There have been a number of claims of detection of mass flows in late-type stars, which however are not exempt of some controversy (e.g., Lim and White 1996). Different methods have been used to attempt direct measurements of mass loss rates in M-type stars, most notably using *mm* wavelength observations (e.g., van den Oord and Doyle 1997). Recently there have been important developments towards indirect detections of stellar winds through their interactions with the surrounding ISM. These observations were performed either in X-rays (Wargelin and Drake 2002) or in the strong Lyman-α emission feature (Wood et al. 2002, 2004, 2005). In particular, recently the stellar mass loss rates and related stellar winds were estimated for several nearby main sequence G- and K-type stars using Hubble Space Telescope (HST) high-resolution spectroscopic observations (Wood et al. 2002) and the analysis of the hydrogen Lyman-α absorption features.

The HST observations by Wood et al. (2002, 2004, 2005) revealed neutral hydrogen absorption features associated with interaction between the stars' fully ionized coronal winds and the partially ionized local ISM. From these absorption features the mass flux of the stellar wind as a function of stellar activity was modeled. Based on the study of a small sample of observed solar-like stars it was concluded that the mass loss and therefore the stellar wind mass flux decreases with stellar activity and stellar age. From these observations one can conclude that younger solar-like stars have much denser and faster stellar winds than the present Sun. A correlation between the mass loss rates and the X-ray surface flux indicates an average solar wind density of 100–1,000 times higher than today during the first hundred Myr after a solar-like star reaches the ZAMS (Wood et al. 2002, 2004, 2005). However, one should not forget that these results follow from observations of only a few K- and G-stars and to get a detailed knowledge of the stellar wind mass flux and mass loss of young stars more observations are needed in the future (Wood et al. 2005).

Furthermore, it is known from observations of our Sun that active stars possess strong eruptions of coronal mass (e.g., CMEs), occurring sporadically and propagating as large-scale plasma-magnetic structures, which disturb the interplanetary space. CMEs appear as one of the important factors of the stellar activity. Traveling outward from the star at high speeds (up to thousands kilometers per second in the case of the Sun), CMEs create major disturbances in the interplanetary medium and produce

Table 4 Power-law approximated minimal and maximal CME plasma densities as function of distance from a Sun-like star (Khodachenko et al. 2007a,b)

d(AU)	$n_{\mathrm{CME(min)}}$ (cm^{-3})	$n_{\mathrm{CME(max)}}$ (cm^{-3})
0.005	$\sim 10^6$	$\sim 10^8$
0.01	$\sim 10^5$	$\sim 10^7$
0.05	$\sim 10^4$	$\sim 10^5$
0.1	$\sim 10^3$	$\sim 10^4$
0.2	∼100	∼500
0.5	∼50	∼70
1.0	∼7	∼10

strong effects on planetary environments and magnetospheres. Because of the relatively short range of propagation of the majority of CMEs, they should most strongly impact the magnetospheres and atmospheres of close orbit (<0.1 AU) planets. This makes the terrestrial exoplanets at orbital locations of close-in HZs around low mass active M-type stars (Khodachenko et al. 2007a,b; Lammer et al. 2007) especially affected by stellar CMEs.

Solar CME plasma density n_{CME} at distances $\leq 30 R_{\mathrm{Sun}} \approx 0.14$ AU is estimated from the analysis of brightness enhancements in the white-light coronagraph (LASCO) images associated with CMEs. At large distances (>0.3 AU) n_{CME} is determined from the in-situ density measurements of magnetic clouds. Based on these data Khodachenko et al. (2007a,b) provided combined power-law approximations of CME density decrease with the distance to a star (Table 4): $n_{\mathrm{ejecta}}^{\mathrm{min}}(\mathrm{d}) = 4.88\ (d/d_0)^{-2.3}$ and $n_{\mathrm{ejecta}}^{\mathrm{max}}(\mathrm{d}) = 7.10\ (d/d_0)^{-3.0}$, where $d_0 = 1$ AU and d is taken in AU. Besides that, the average mass of solar CMEs was estimated to $\approx 10^{15}$g, whereas their average duration at distances (6–10) R_{Sun} is close to ≈ 8 h.

Table 4 shows the estimated minimum and maximum CME proton densities as function of orbital distance. Until now, indications for stellar CME activity have mainly come from absorption features in the UV range during the impulsive phase of strong flare events, or from the blue-shifted components in time series spectra. Such features were found in the spectra of several active M-type stars, like EV Lac (Ambruster et al. 1986), AD Leo (Houdebine et al. 1990), and AU Mic (Cully et al. 1994) and were interpreted as indications of rapid mass ejections.

In general, Cully et al. (1994) concluded that CMEs on those stars might be much stronger than the solar events. For example, Large enhancements in the far blue wings of the Hγ and Hδ lines during a strong flare on AD Leo, observed by Houdebine et al. (1990) were explained as a powerful CME-event with an estimated line-of-sight speed 5,800 km/s; 7.7×10^{14} kg mass and 5.0×10^{27} J kinetic energy. Whereas, modeling of the observed characteristics of a flare on AU Mic, detected with the Extreme Ultraviolet Explorer have led Cully et al. (1994) to propose a plasmid with the initial ejection speed of 1,200 km/s and mass 10^{17} kg. Discrete absorption features typical for a CME were also found in UV spectra of the eclipsing binary system V471 Tauri, consisting of a K-type dwarf star and a white dwarf.

These features were interpreted by Mullan et al. (1989) as the signatures of a material cloud, ejected from to the K-dwarf and moving relative to it with velocities of 700–800 km/s, which are comparable to the solar CME speeds. Bond et al. (2001), analyzing the absorption features in HST spectra of V471 Tau, estimated the frequency of CME-events of the K-type star component of the system as 100–500 CMEs per day, which is approximately 100 times more than the CME occurrence frequency of the present Sun.

Khodachenko et al. (2007a) provided an estimation of the critical stellar CME production rate (CME occurrence frequency), for which (and higher) a close orbit planet will experience a continuous action of the stellar CMEs, so that the discrete collisions of a planet with CMEs can be replaced by a continuous action of the CME plasma flow. This corresponds to a situation when each CME collides with the planet during the time interval of the previous CME passage over the planet. The estimates gave the critical CME production rate of about 36 CME per day, which is not very much higher than the CME occurrence frequency of the present Sun ($\geq$6 CME per day). Under the conditions of continuous CME flow action on a planet, the parameters of the stellar wind (density, speed, etc.) should be replaced by the parameters of the CME plasma, which will mean harder and more extreme conditions for the planetary environments than that in the case of a regular stellar wind.

To summarize this section we would like to emphasize that the HZs of active flaring stars are exposed to intense stellar XUV irradiation, extreme particle and stellar wind conditions over rather long time periods. This radiation and particle exposure would strongly impact the atmospheres of terrestrial type planets in these HZs and may therefore significantly limit their range, as compared to those that result from the traditional HZ definition, which is based on the pure climatological approach.

4 Habitability relevant factors which are not considered in the classical habitable zone definition

4.1 Basic geophysical conditions

Besides liquid water on the surface of a planet, a second characteristic of habitable planets is an atmosphere which is dense enough that it can stabilize the surface temperature of the planet through climate feedback like the "greenhouse effect" (e.g., Kasting et al. 1993). The greenhouse effect is caused by compounds that are very efficient absorbers in the infrared but not in the visible. The visible light of the parent star reaches the planetary surface and is remitted in the infrared where part of the energy is absorbed by the atmospheric greenhouse gases, increasing the temperature of the planet. Carbon dioxide (CO_2), methane (CH_4) and water (H_2O) raise the surface temperature on Earth by an average of 15°C, above the freezing point freezing. A planet has to accrete enough volatiles during its formation to have an atmosphere and it has to be massive enough to maintain this atmosphere.

Venus and Mars demonstrate the limits of planetary habitability for life as we know it on Earth. Venus is the best example of what happens to a planet when the surface temperature exceeds a certain limit (like at the inner edge of the HZ). Venus has a surface pressure of about 90 bars and a surface temperature of about 480°. Present Venus has an atmosphere but only a tiny amount of water. It is uncertain if it did have water after its formation. If Venus had a similar water reservoir as Earth it was converted to vapor due to the high surface temperatures of the planet. It remains an open question whether Venus lost its water due to a "runaway greenhouse" (Ingersoll 1969; Rasool and deBergh 1970; Walker et al. 1970) or a "moist greenhouse" (Kasting 1988, 1992) and thermal and non-thermal escape processes triggered by the young active Sun (Kulikov et al. 2006, 2007; Lammer et al. 2008).

The "runaway greenhouse" occurs when water vapor increases the greenhouse effect, which in turn increases the surface temperature, leading to more water vapor that in turn heats the atmosphere. The other scenario is the "moist greenhouse" where water is lost once the stratosphere becomes wet but most of the water of the planet remains liquid. For both cases the loss of water happens high up in the atmosphere where water is photolyzed and H_2 escapes while oxygen reacts with the planetary crust or is lost to space via non-thermal escape processes (Kulikov et al. 2006, 2007).

Today Mars is a dry, frozen desert that cannot sustain life on its surface. The Martian atmosphere was lost and became too thin to warm the planetary surface. A minimum planetary size for a terrestrial planet seems to be a crucial factor for an Earth-analog habitat. About >4–4.5 Gyrs ago Mars may have had an atmosphere thick enough to maintain liquid water on the surface (e.g., Kulikov et al. 2007 and references therein) which led to the hypothesis of life on early Mars. Large impactors evaporated the Martian atmosphere and the low gravity of the planet was not able to retain the gas in the hot plumes created by those impactors (Melosh and Vickery 1989; Manning et al. 2006; Pham et al. 2009).

4.2 Planets with and without plate tectonics

The most plausible model is that small bodies very quickly develop a stable thick lithosphere within the first 100 Myr (e.g., Nelson 2004), resulting in a smaller and weaker heat flow and therefore smaller convection cells. Liquid magma is still present within smaller bodies, which is shown by the fact that the maria at the Moon caused by large impacts are covered by lava after the late heavy bombardment (LHB) and, e.g., Olympus Mons on Mars.

No other planetary body in the Solar System than Earth seems to have developed a system of continents or cratons, light-weight parts of the crust which do not vanish by subduction. Sediments are produced, which result in the first gneisses and granites, probably already before the LHB. The oldest formations are dated to about 4 Gyr ago, but some zircons are dated earlier and must have been formed about 4.4 Gyr ago (Nelson 2004). If the separation of continental crust on Earth started so early, all other terrestrial bodies should be expected to have experienced such a phase too. If started, plate tectonics will produce continental crust at subduction zones which are added to the cratons over time. Another model postulates that continents have been in a balance of accretion and constructions since 4 Gyr (Arndt 2004). Probably there are three stages:

– a magma ocean 4.5–4.0 Gyr ago,
– a transition phase with starting plate tectonics but heavy crustal breaks by superplumes 4.0–3.0 Gyr ago,

– since then the development of slow plate tectonics up to the present velocities (Eriksson and Catuneanu 2004).

It seems that this process producing terranes, increasing continents by lateral accretion of pieces of the crust, and formatting of supercontinents by driving cratons together, has a period of a few 100 Myr (on Earth Vaalbara, Kenorland, Columbia, Rodinia, Pangaea). If there were plate tectonics on all terrestrial bodies, it seems likely that it stopped very early on small bodies, at least about 3.5 Gyr ago, because the LHB impacts are still seen as mentioned before. On Earth the first signs of life are detectable at that time (e.g., Schopf 2004).

The picture at Venus seems to be more complicated. The two bodies of Earth's size in the Solar System, Venus and Earth, differ heavily in the number of impact craters. The approximately 100 craters at Earth are much less than the approximately 900 on Venus. Even if some on Earth vanished by erosion processes this means that tectonic processes extinguished most of the craters on Earth, but not on Venus. The question is why Venus did not develop active plate tectonics similar to the Earth? One factor might be the high surface temperature which weakens the heat flow and slows down the convection (Valencia et al. 2007).

Despite about 900 craters (http://www.lpi.usra.edu/resources/vc/vchome.shtml) from impacts, the landscape of Venus is dominated by lava flows and probably other formation processes. The sharp decline of Ishtar Terra to the West and the much softer one to the East may point to a subduction zone in the West caused by plate tectonics (Ansan et al. 1996; Lenardic et al. 1991; Vorder Bruegge and Head 1990; Janle and Jannsen 1984). In other regions plumes and superplumes might form large igneous provinces (LIP), providing heat transfer and producing surface structures (Nijman and de Vries 2004) like plateaus and volcanoes (hot spots). Plate tectonics should have stopped at Venus after the water inventory was lost (Nelson 2004). Without any surface probes the geological history of Venus cannot be deciphered.

Plate tectonics together with permanent liquid water most likely provided the first environment for the evolution of life, the black smokers. Plate tectonics does not only regulate the composition of a terrestrial atmosphere by the cycling of volatiles, including the greenhouse gas CO_2 and hence the surface temperature and planet habitability (e.g., Sundquist 1993; Kasting et al. 1993; Franck et al. 2000; Wolstencroft and Raven 2002), but it is driving evolution by always changing the environment. The chances for being created and being forced to evolve (and probably also for being destroyed) are better on an active planet than on a stable one.

As illustrated in Fig. 4 plate tectonics is also an important factor for the generation of an Earth-like long-time strong intrinsic planetary magnetic dynamo, which protects the atmosphere from solar wind erosion and deflects high energy cosmic rays (e.g., Ward and Brownlee 2000; Grießmeier et al. 2005; Khodachenko et al. 2007b). Although the driving mechanisms for plate tectonics are not fully understood, the minimum requirements are a sufficient mass relevant for the heat flow to drive mantle convection, and water to lubricate plate motion (e.g., Regenauer-Lieb et al. 2001; Solomatov 2004). Water is the lubricant that allows the plates of the crust to slide and subduct, without water in the mantle the evolution of the planetary mantle and planetary tectonic engine would stop. Water makes the lithosphere deformable enough for

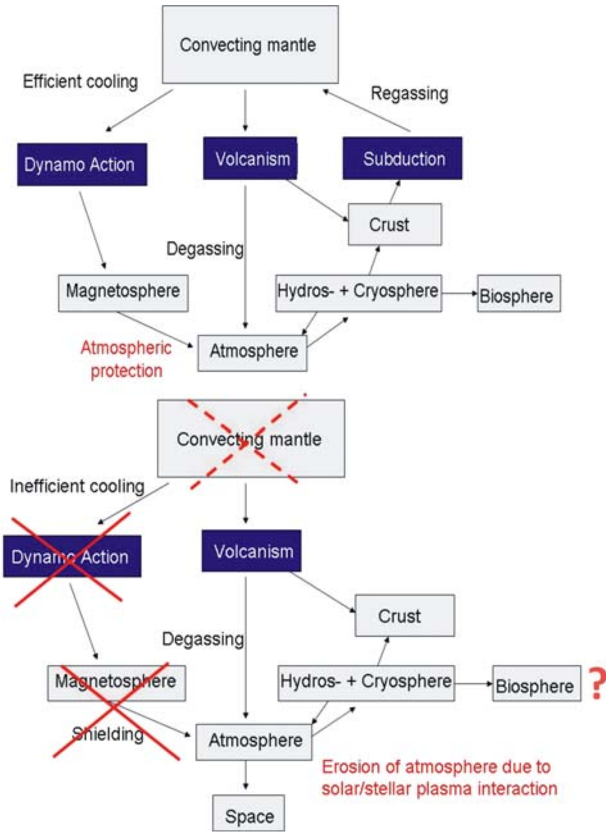


Fig. 4 Illustration of the connection between plate tectonics, mantle cooling and magnetic Earth-type dynamos. *Upper panel* Active geodynamic conditions due to plate tectonics. *Lower panel* Inefficient cooling in the mantle yields a one-plate planet with non or weak magnetic protection (courtesy of D. Breuer)

subduction of the crust to occur and it reduces the activation energy for creeping and the solidus temperature of mantle rock, thereby enhancing the cooling of the interior and the efficiency of volcanic activity.

Large water reservoirs in the mantle and on the surface are interacting. The mantle loses water and other volatiles like CO_2 through volcanic activity, and therefore helps to sustain the atmosphere. On the other hand, water and CO_2 are recycled together with the subducting crustal rock (e.g., Breuer et al. 1996, 1997). Furthermore, the recycling of the crust through plate tectonics keeps the crust thin, which seems to be mandatory for plate tectonics to operate. If the crust is too thick, the lithospheric plate comprising the crust will be too buoyant to be subducted because on Earth the cratons became never subducted anymore after formation. Finally, plate tectonics seem to help to establish the right temperature conditions in the interior that are required to maintain the action of a strong magnetic dynamo for several billion years.

For terrestrial exoplanets the conclusion can be drawn that planets with sizes less than Earth or Venus (e.g., Martian sized bodies) lose their ability for plate tectonics very quickly. Water-rich Earth-sized bodies should maintain the convection cycle for a long time, thus forming dry continents and basins with water. Concerning "super-Earths" which are terrestrial exoplanets with sizes between $R_{\oplus}$ and 2 $R_{\oplus}$ the kinematic energy of the convection cells should be higher than on Earth. Plate velocities should be faster and orogenesis much heavier, resulting in higher mountains and more volcanic activity. It was shown by Valencia et al. (2007) that the crust on "super-Earths" which is comparable to the "oceanic crust" on Earth is even slightly thinner and the convection cells are only slightly larger, but plate velocities will increase approximately linearly with planet mass. However, the conclusion drawn from this, namely that mountains will not be higher and trenches not deeper seems to be questionable (Aguilar 2008). A faster plate velocity will result in a higher kinematic energy which produces orogenesis at subduction and collision zones. Plate tectonics on "super-Earths" depends on several factors, like the original water contents, the forming of cratons and the distribution of convection cells, all probably individual for each planet. Even on Earth variations are frequent within space and time and it is difficult to find a precise model to describe the processes leading from the magma ocean to the present day continents.

4.3 Magnetic dynamos and their role in atmospheric protection against solar/stellar plasmas

Strong intrinsic planetary magnetic fields are generated by a complex interaction of hydromagnetic processes. The source of the internal magnetic field is the motion of a highly conductive fluid within the planet (i.e., the liquid outer core for terrestrial planets, or a region of electrically conducting hydrogen for gas giants). These large-scale motions can be produced by two different mechanisms that are believed to appear subsequently during the evolution of a terrestrial planet (Stevenson 1983, 2003; Stevenson et al. 1983), namely thermal and compositional convection.

Large amounts of water and related plate tectonics could have played a role in keeping a magnetic dynamo alive over geological time scales. Depending on the initial water inventory of early Venus, possible plate tectonic activity and the related thermal history of Venus' interior, theories of dynamo generation suggest that there could have been a magnetic moment on Venus of the same order as that of present Earth for about the first billion years after the origin of the planet (Stevenson et al. 1983). During that time, thermal convection from the heat left over from accretion could have driven the dynamo. After that this energy source diminished and no other source replaced it. The solid core formation in Earth's interior keeps its dynamo working to this day by virtue of the related stirring of the molten layer around it. Present Venus may lack the necessary internal chemical or physical ingredients for solid core formation, or such processes may have stopped at an earlier time (Stevenson et al. 1983).

Because the solidification of the inner core is thought to be the energy source for the present day terrestrial magnetic field, and smaller bodies thermally evolve more rapidly than larger bodies, one can conclude that the terrestrial planets today are in three different magnetic phases, where Venus is most likely in a predynamo phase, not having cooled to the point of core solidification (Russel 1993). The Earth is in a dynamo phase and Mars is in a post-dynamo phase. Maintaining an intrinsic magnetic moment would have been problematic following the loss of water.

Present Mars lacks a detectable global magnetic field (Acuña et al. 1998, 2001; Connerney et al. 2001). On the other hand, at small spatial scales the Martian crust shows remnant magnetism, which is, on average, about 10 times more magnetized than that of the Earth (Connerney et al. 2004). This suggests that Mars once had an active dynamo and the crustal remnants were acquired either by thermoremnant magnetization or by another process like chemical remnant magnetization. Acuña et al. (2001) proposed that Mars' magnetic dynamo ceased operation relatively early in the planet's evolution. These authors argued that large areas above the huge impact basins—Hellas, Argyre, Isidis lack measurable crustal fields, as did much of the impact basins in the northern lowlands. If these impact basins had originated in the presence of a strong ambient magnetic field, they would most likely have been magnetized as the rocks cooled below the Curie temperature, because a reworking process or shock demagnetization of the entire crust in the vicinity of these impact basins is very unlikely. Since the formation of these large impact basins is believed to have occurred at the end of the LHB ∼3.8 Gyr ago, one can conclude that the Martian dynamo ceased its operation after a few hundred Myr during the Noachian epoch. This hypothesis is in agreement with the model of Breuer and Spohn (2003) which predicts thermal convection early in the Martian history, so that the dynamo action could occur for hundreds of Myr.

On the other hand Schubert et al. (2000) and Stevenson (2001) proposed that these large impact basins originated before the onset of the dynamo, because the magnetization of the southern highlands could also result from localized heating and cooling events that occurred after large impacts and basin formation. Further, Schubert et al. (2000) note that an early onset and cessation of the Martian dynamo is difficult to reconcile with the hypothesis of a dynamo driven by solidification of an inner core as it is thought to have occurred on Earth. Phase transitions in the Martian mantle and their interaction with the mantle flow may provide an alternative explanation for a late onset of a purely thermal dynamo (Spohn et al. 1998). They discussed how a putative layer of the mineral perovskite above a small and Fe-rich core could affect the thermal history of the core evolution. According to 2-D and 3-D convection model calculations by Breuer et al. (1997), a 100–200 km thick perovskite layer reduces the heat flow from the core since it forms an additional convective layer between the overlaying mantle and the core. A perovskite layer (Spohn et al. 1998) could thus be responsible for a late onset of the Martian dynamo, or a re-birth of the dynamo at a later time (Lillis et al. 2006).

Thus, the models of the core evolution are consistent with both an early magnetic field and or a late onset of it (Connerney et al. 2004; Lillis et al. 2006). However, these models cannot predict the strength of the magnetic field. Unfortunately, there is as yet no observational constraint on the timing of the Martian dynamo. An early Martian dynamo with a minimum and maximum expected magnetic moment between 0.1 and 10 times that of the present Earth can be expected (Schubert and Spohn 1990).

4.3.1 The role of rotation and magnetic dynamos

As discussed above, the magnetic field protects a planetary atmosphere against strong erosion by the solar or stellar wind and CMEs. Planets that are not protected by sufficiently strong magnetic shields and exposed to extreme stellar radiation and plasma flows can be rendered inhabitable due to non-thermal atmospheric erosion processes. Because planetary rotation and magnetic moment are linked, rotation is probably an important factor for planetary habitability. This is especially relevant for terrestrial planets in the HZ around M dwarfs. These planets are located so close to the star that they are tidally locked, i.e., have strongly reduced rotation rates (e.g., Grießmeier et al. 2005, 2008; Khodachenko et al. 2007b; Lammer et al. 2007). In many cases, the corresponding weak magnetic moment may not be sufficient to protect the atmosphere, leading to planets which may lose their atmosphere and water inventories even when they are located within the liquid water HZ.

The generation of planetary magnetic moments is based on convective motion in the planetary core. As a requirement for convection to occur, the Coriolis force has to have a large effect on the flow. This condition, however, is easily satisfied, even for slow rotators like Venus (Stevenson 1983, 2003). Thus, when we discuss the influence of rotation on the magnetic dynamo further below, we always assume that rotation is strong enough for a dynamo to exist in the first place.

4.3.2 Dynamo models with rotational influence

Under the assumption that the planetary magnetic field can be derived from a dipole moment (i.e., no higher multipoles), one can attempt to estimate the planetary magnetic dipole moment M from characteristic values of the planet using simple analytical models. This usually leads to simple scaling relations, which can be used to calculate the order of magnitude of the planetary magnetic dipole moment (or its surface magnetic field, which is proportional to Mr^{-3}, where r is the planetary radius). Several such models are summarized in Grießmeier et al. (2005). All these models yield increasing magnetic moments with increasing planetary rotation rates. Thus, the shorter the planetary rotation period is (i.e., the faster its rotation), the larger the resulting magnetic moment can be expected to be.

4.3.3 Dynamo models without rotational influence

While the analytical models considered above predict an increase of magnetic moment with increasing rotation rate, numerical experiments indicate that the magnetic moment may be independent of the angular frequency. Christensen and Aubert (2006) and Olson and Christensen (2006) have studied numerically an extensive set of dynamo models in rotating spherical shells, varying all relevant control parameters by at least two orders of magnitude. These simulations indicate that the magnetic field is basically controlled by the buoyancy flux, and rotation does not seem to have any influence. This result is in contradiction with the analytical models discussed above.

To estimate the maximum impact on an atmosphere we assume that rotation has an influence on the planetary magnetic moment. Further studies are needed to clarify the relation between planetary rotation and magnetic field strength.

4.3.4 Estimation of magnetic moments for terrestrial planets with different size, mass and rotation periods

Based on the formulas given by Grießmeier et al. (2005) one can calculate the magnetic moments for different situations. To illustrating the possible effect of the rotation on the strength of a magnetic dynamo we investigate a "super-Earth" and an "Earth-like" planet. We assume the following planetary structure (Cain et al. 1995; Léger et al. 2004):

– the radius for the massive terrestrial planet of 6 $M_\oplus$ is about 1.63 $r_\oplus$,
– the core density is 10.6 10^3 kg/m^3 for the Earth-like, and 15.5 10^3 kg/m^3 for the large/massive planet case,
– the size of the planetary core is $r_c = 0.55\ r_\oplus$ for the Earth-like, and $r_c = 0.52\ r_\oplus$ for the large terrestrial planet case

and

– the conductivity σ is roughly the same as on Earth for both cases.

The resulting magnetic moment as a function of the planetary rotation period is given in units of the Earth's current magnetic moment ($\mathcal{M}_\oplus = 8\ 10^{22}$ Am^2).

The shaded areas in Fig. 5 represent the ranges of results obtained by the different magnetic moment scaling laws. For a rotation period of 10 h or more, one finds that Earth-like terrestrial planets can have a maximum magnetic moment of 2.5 $\mathcal{M}_\oplus$. Tidally locked planets have rotation periods equal to their orbital period (i.e., rotation periods 10–100 times larger than for the Earth), and are likely to have a much lower magnetic moment than the Earth (more than one order of magnitude lower).

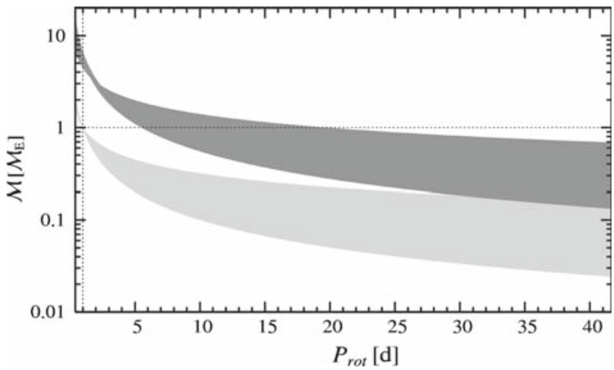


Fig. 5 Ranges of expected planetary magnetic dipole moments as a function of rotation period and planetary mass. *Light grey*: Earth-like exoplanets. *Dark grey*: Super-Earths with 6 Earth masses and 1.63 Earth radii (Léger et al. 2004). *Dotted lines* indicate the current value of the Earth (rotation period and magnetic moment)

Large terrestrial planets of 6 Earth masses are likely to have a magnetic moment approximately a factor of 5 higher than Earth-like planets. As the planetary radius is also larger (1.63 $r_{\oplus}$), this corresponds to a surface magnetic field approximately 20–30% higher than on an Earth-like planet. For this reason, the magnetic shielding of a large planet is slightly better than that of an Earth-sized planet (assuming identical rotation rates). The improved magnetic shielding with respect to atmospheric erosion is presented in Sect. 4.3.5. The temporal evolution of the magnetic moment of an Earth-like planet caused by the temporal evolution of the rotation rate is discussed by Dehant et al. (2007).

4.3.5 Magnetic protection of slow rotating planets

Planets with a small magnetic moment also have a small magnetosphere. This is quantitatively discussed for terrestrial exoplanets in Grießmeier et al. (2005, 2008) and Khodachenko et al. (2007b).

For a small magnetic moment and/or a strong enough plasma flow (either stellar wind particles or solar like stellar CMEs), the magnetosphere is so small that its boundary surface (the magnetopause) is located below the outermost layers of the planetary atmosphere (Khodachenko et al. 2007b; Lammer et al. 2007). The results shown in Fig. 6 indicate that one can expect many planets within the HZ of dwarf stars with Venus- or Mars-like stellar plasma-atmosphere interactions.

However, we have to keep in mind that dynamos and the resulting strong intrinsic magnetic fields, as on Earth, are only generated if the planet is geophysically active. If plate tectonics stop or do not work well on the planet, the magnetic dynamo would also stop working. As discussed in Sects. 4.2 and 4.3, as a result one obtains Venus- or Mars-like unprotected planetary environments where the ionized particles within the upper atmosphere form a planetary obstacle.

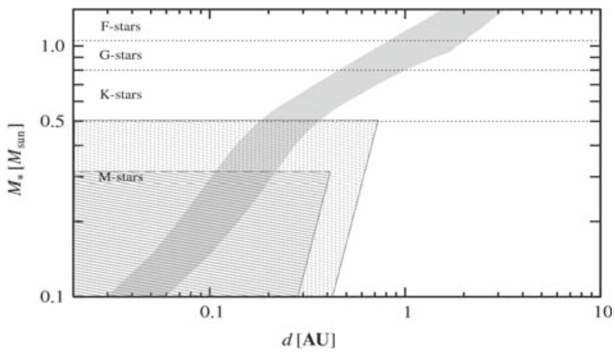


Fig. 6 Region where, according to the estimations of Sect. 4.3.4, the magnetopause lies close to the planetary surface. *Grey-shaded area* the classical liquid water HZ as discussed in Sect. 1 (e.g., Kasting et al. 1993). *Lightly dotted area*: region where the magnetopause can be compressed to altitudes <1,000 km above the planetary surface by strong stellar CMEs for an Earth-like planet. *Darker dotted area* same for a super-Earth with 6 times the mass of the Earth and 1.63 times its radius (Léger et al. 2004). It can be seen that even if the surface magnetic field is only slightly higher, as indicated above, larger planets are better protected against atmospheric erosion by strong CMEs due to their expected larger magnetic fields

4.4 The role of ionospheres and induced magnetic fields in atmospheric evolution

Because low mass stars are very numerous compared to Sun-like G-type stars, one can expect that there may be many slow rotating terrestrial planets within close-in HZs. As pointed out before, due to the slower rotation these planets should have weaker magnetic fields so that the stellar plasma interaction resembles Venus rather than Earth. In such cases the stellar plasma flow will interact directly with the ionized part (ionopause) of the upper atmosphere of the exposed planet. The ionosphere also plays an important role in controlling the atmospheric loss from Earth-type planets. All planets with substantial atmospheres (e.g., Earth, Venus, Mars, and Titan) have ionospheres which expand above the exobase. Atmospheric escape related to the ionosphere depends non-linearly on the solar/stellar EUV/FUV conditions and solar/stellar wind condition in a complicated manner. By synthesizing observations from the Earth, Venus, Mars, and Titan, we can qualitatively evaluate how the basic controlling parameters affect the various loss mechanisms through the ionosphere for both magnetized and unmagnetized planets illustrated in Fig. 7.

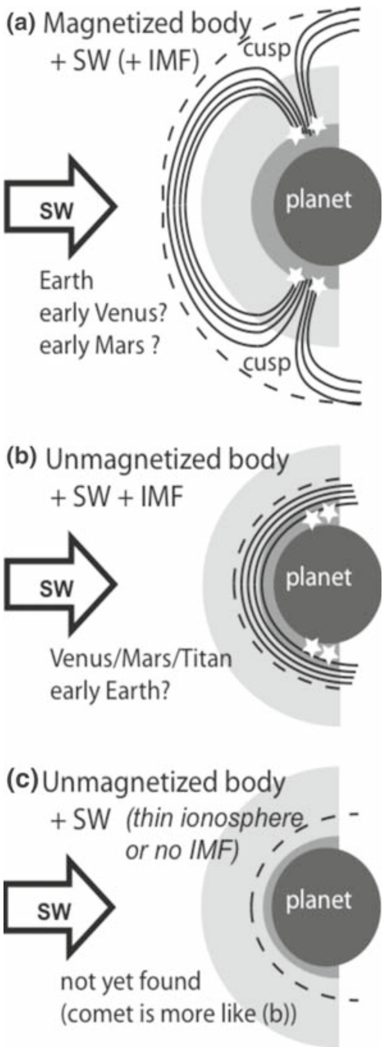
The black, dark grey, and light grey areas in Fig. 7 denote the solid planet, ionosphere, and exosphere, respectively. The solid lines denote the magnetic field, and the dashed line denotes the shielding boundary from the shocked solar wind, respectively. The white stars denote the region where non-thermal (electromagnetic) ion heating is observed in the dayside ionosphere at both Earth and Mars.

For a magnetized planet (Fig. 7a), the solar/stellar plasma is deflected by the magnetopause where the solar wind dynamic pressure (SWDP) is balanced with magnetic pressure of the planetary magnetic field, except in singular regions that are called the "cusps". Through the cusps, the solar/stellar plasma and energy directly and continuously reach the ionosphere (Heikkila and Winningham 1971; Yamauchi and Lundin 2001; and references therein). This localized energy generates strong electromagnetic wave activities in and above the ionosphere, which results in wave-related non-thermal ion heating and escape around the cusp (e.g., Moore et al. 1999; and references therein). For an unmagnetized or weakly magnetized planet (Figs. 7b, c), the solar/stellar plasma flow is blocked by the ionopause, a magnetically shielding boundary the magnetic pressure of which is balanced by the SWDP outside, and by the ionospheric plasma pressure inside, respectively, in a collision-free regime. The ionopause magnetic field is created by the ionospheric induction current through the "pile-up" mechanism (e.g., Alfvén and Fälthammar 1963; Russell et al. 2006).

Because the frequent change of the interplanetary magnetic field (IMF) direction always rearranges this shielding magnetic field, the ionopause boundary is less stable than the magnetopause, allowing more frequent localized energy deposition than through the magnetopause. Such a localized energy deposition aids to non-thermal ion heating and escape. The ionosphere contributes directly to atmospheric loss processes in various ways (Yamauchi and Wahlund 2007):

– It is a source of non-thermal ion escape by, for example, non-thermal ion heating for both magnetized and unmagnetized planets. The resulting escape has a higher O/H ratio than the thermal escape or ion pickup loss, which is important

Fig. 7 Illustration of the three known solar wind plasma interaction conditions with planetary bodies (from Yamauchi and Wahlund 2007)



for determining the oxidization state of the atmosphere, and is more SW/radiation dependent than thermal escape, according to observations.

– It is a protector against a solar wind-driven escape of ions by gyro-trapping to the piled-up IMF for unmagnetized planets. Therefore, the ionopause location relative to the exospheric extent is a controlling factor for the ion-pickup loss. Ionization of neutrals also reduces escape because ions are bound by the ionopause magnetic field. For a magnetized planet, the pick-up loss of the exospheric neutral atoms is very small because the exosphere does not extend (much) above the magnetopause.
– It also contributes to plasma instabilities that lead to large-scale momentum transfer and related escape. In addition, the mass loading effect by the ionospheric ions enhances large-scale SW interaction with the magnetosphere/ionosphere for both

magnetized and unmagnetized planets. As a result, the ionosphere and its O/H ratio make the solar parameter dependencies of the atmospheric loss very dynamic for both thermal and non-thermal escape mechanisms.

4.4.1 The ionosphere as a source of particle loss

The observed non-thermal mass loss rates from present Earth are summarized in Table 5 from the present terrestrial ionosphere during solar maximum (e.g., Moore et al. 1999; Cully et al. 2003). Note that the thermal (Jeans) escape of hydrogen is estimated (no direct measurement exists) to be smaller than this non-thermal escape of protons during solar maximum (Vidal-Madjar 1978; Kasting and Catling 2003, and references therein).

For reference, the observed non-thermal escape rates from present Mars are about 0.1–1 kg/s for heavy ions (Lundin et al. 1990) and 0.02–0.1 kg/s for picked-up protons (Yamauchi et al. 2006), and about 1–2 kg/s for heavy ions from Venus (Brace et al. 1982; Lundin et al. 2007).

The observed terrestrial non-thermal ion escape of the heavy ions (O^+ and N^+) from the ionosphere increases much more than escape of H^+ during high solar activity (which correlates with high $F_{10.7}$ flux) and during major magnetic storms (Chappell et al. 1982; Hamilton et al. 1986; Cully et al. 2003). When the $F_{10.7}$ flux increases by a factor of 3, i.e., from solar minimum to solar maximum, the O^+ outflow increases by a factor of 100 while the H^+ outflow increases only by a factor of 3. By extrapolating this result one can see that an increase of the $F_{10.7}$ flux by an order of magnitude (corresponding to the early Sun or an active dwarf star) might cause an increase of the non-thermal ion escape from the ionosphere in a range from four orders of magnitude (for power law extrapolation) to nine orders of magnitude (for exponential extrapolation) for O^+ and by 1–2 orders of magnitude for H^+ ions.

The O/H ratio of escape rates from an early planet/exoplanet is expected to be high enough to keep the atmosphere reduced (H_2-rich) rather than oxidized. The non-thermal escape flux of ions from Mars was recently found to be positively correlated with the SWDP (Lundin et al. 2007) and enhanced during solar energetic particle events (Futaana et al. 2008). For Venus the solar parameter dependence of the non-thermal ion escape is not clear (Mihalov et al. 1995). It is only known that variable IMF increases loss by detached ionospheric clouds, though Brace et al. (1982) found positive dependence of the ionotail density on the solar EUV/FUV flux and a negative dependence on the SWDP.

Table 5 Budget of non-thermal ions above the Earth's ionosphere

Species	H^+ (kg/s)	O^+ (kg/s)	Meteor (kg/s)
Out ($h = 2$–$4\,R_\oplus$)[a]	0.03–0.15	0.3–5	–
In ($h = 840$ km)	0.003–0.02	?	0.5

Range is between geomagnetic quiet time and active time
[a] DE-1 data during solar maximum

4.4.2 The ionosphere as an atmospheric protector

A high EUV/FUV flux generates a hot and expanded exosphere which facilitates enhanced escape. However, a high EUV/FUV flux also causes a hot and expanded ionosphere. Earth, Mars and Venus observations show large variation of the ionospheric density and its extent (or the ionopause height) from solar minimum to solar maximum (Evans 1977; Zhang et al. 1990; Kliore and Luhmann 1991).

The expanded ionosphere reduces the ion pick-up loss because only the neutrals above the ionopause can contribute to the pick-up process. The expanded ionosphere may also reduce the enhancement in thermal escape by ionizing a portion of neutrals with escaping velocity inside the ionosphere where ionizing electron density is high. At present we do not know the amount of the reduction and hence whether the total pick-up loss increases or decreases for an enhanced EUV/FUV flux.

The ionopause location is controlled by the SWDP and little by the IMF, both of which are highly variable. An increase in the SWDP results in a decrease of the pressure balance altitude, which moves toward the planet. In case the exosphere temperature does not vary with the SWDP the shrunk ionopause means more neutrals existing beyond the ionopause, causing an increase in the pick-up loss. We now consider only the unmagnetized planets as shown in Fig. 7b, c.

4.4.3 The ionosphere and large-scale momentum transfer

The magnetopause for a magnetized planet and the ionopause for an unmagnetized planet are known to be the main source regions for a large-scale plasma instability that leads to the large-scale momentum transfer. The instability is enhanced by injection of heavy ions into the region (magnetopause case) or by simple expansion of the unstable region (ionopause case).

An increase in the EUV/FUV flux causes expansion of the ionopause, which increases the large-scale momentum transfer for an unmagnetized planet, while it increases the number of ionospheric ions and the relevant instability in the magnetopause for a magnetized planet.

The mass loading-related interaction increases when the EUV/FUV flux increases for both magnetized and unmagnetized planets. However, the IMF intensity has a minor influence on the ionopause activity because the ionopause is mainly determined by the SWDP and the ionospheric pressure. Tables 6 and 7 summarize the expected solar parameter dependences of the ion escape rates for magnetized and unmagnetized planets.

A high loss rate is expected when the planet is small, the SWDP is high, the IMF is variable, and the EUV/FUV flux is high. However, the dependence is different for the thermal escape (small change) and non-thermal escape (large change). It might even be possible that the EUV/FUV dependence of the pick-up loss is negative.

Since the O/H ratio of the observed non-thermal escape drastically increases on for a high EUV/FUV, SWDP, IMF or a variable IMF, the ancient Earth and other planets/exoplanets must have had a much higher O/H ratio of atmospheric escape rates

Table 6 Major non-thermal ion escape mechanisms from magnetized planets

Increase in	Pick-up (small)	Large-scale momentum transfer	Non-thermal heating of ions	cf. Jeans and photo-chemical	O^+/H^+ ratio of escape
EUV/FUV flux	No change	[a]	Large increase	Increase	[b]
SWDP	Some increase	increase	Large increase	No change	Large increase
IMF intensity	No change	[a]	Increase	No change	Increase
IMF variability	No change	[a]	Increase	No change	Increase
Gravity	Some decrease	No change	?	Some decrease	Decrease

SWDP solar wind dynamic pressure, *IMF* interplanetary magnetic field
[a] Activity itself increases but the amount of loss is an open issue
[b] Increase or decrease depending on the relative importance of non-thermal heating

Table 7 Major non-thermal ion escape mechanisms from unmagnetized planets

Increase in	Pick-up (small)	Large-scale momentum transfer	Non-thermal heating of ions	cf. Jeans and photo-chemical	O^+/H^+ ratio of escape
EUV/FUV flux	Increase?[a]	Increase	Increase	Increase	[b]
SWDP	Increase	Some increase	Some increase	No change	Large increase
IMF intensity	No change	No change	No change	No change	No change
IMF variability	Some increase	Large increase	Increase	No change	Increase
Gravity	Decrease	No change	?	Decrease	Decrease

[a] Not clear. The comparison between Titan and Mars suggests decrease, but this simple comparison might not appropriate to model the ancient escape
[b] Increase or decrease depending on the relative importance of non-thermal heating

than predicted by models based on current solar conditions. The higher O/H ratio of escape is important for modeling the oxidizing state of the ancient atmosphere.

4.5 Solar/stellar forcing and its impact on atmospheric evolution

As discussed in Sect. 3, a very important factor affecting the atmospheric evolution and planetary water inventories of Earth-like exoplanets is the activity associated with the chromospheres and coronae of their host stars. The relevant physical phenomena include intermittent and energetic flares, CMEs, stellar cosmic rays, enhanced coronal X-rays, and increased chromospheric UV emission (Lundin et al. 2007).

As shown in Table 8 the soft X-ray and EUV radiation of an active star is responsible for heating and expansion of the upper atmospheres (Kulikov et al. 2006, 2007; Lammer et al. 2008; Tian et al. 2008). One can see that the exobase level which separates the collision dominated region of the atmosphere below it and a region above where no collisions occur, moves to much larger distances in an Earth-like N_2-rich atmosphere compared to a CO_2-rich Venus-type atmosphere. High thermospheric temperatures and the related expansion of the thermosphere can result in all light atoms overcoming their gravitational binding to the planet and escaping to space.

The reason for the different behavior of this thermospheric expansion at the planets of more or less similar mass and size is the different cooling rates by IR-radiating molecules in their thermospheres. The most important heating and cooling processes in the upper terrestrial atmosphere can be summarized as:

– heating due to N_2, O_2, and O photoionization by the solar or stellar radiation with $\lambda < 102.7$ nm;
– heating due to O_2 and O_3 photodissociation by the solar or stellar UV-radiation;
– neutral gas heat conduction;
– chemical heating in exothermic reactions;
– IR-cooling in the vibrational-rotational bands of CO_2, NO, O_3, OH, NO^+, $^{14}N^{15}N$, CO, H_3^+, etc.;
– heating and cooling due to the contraction and expansion of the thermosphere;
– turbulent energy dissipation and heat conduction.

On present Earth the thermosphere has an average dayside exobase temperature of about 1,000 K. In the CO_2-rich Venus atmosphere which is closer to the Sun, the exobase temperature on the dayside for average solar activity is about 270 K. The main explanation for this difference is the very efficient cooling by IR emission in the 15 µm CO_2 band.

As shown in Table 8, the exobase on a CO_2-rich Venus-like planet for extremely high EUV fluxes is much closer to the planetary surface and therefore its atmosphere could be better protected against solar or stellar plasma interaction and the resulting atmospheric erosion if the planet had an intrinsic magnetic field.

From Table 8 one can also see that, depending on the solar or stellar EUV flux and planetary and atmospheric parameters, the exosphere could expand even beyond the present Earth's magnetopause which is located at about 10 Earth radii (Roelof and Sibeck 1993; Shue et al. 1997, 1998). In that case, the constituents beyond the magnetopause could be ionized and eroded by the solar wind plasma flow. During its first 500 Myr after the Sun arrived at the Zero-Age-Main-Sequence (ZAMS) the early Earth may have had a higher amount of CO_2 in its thermosphere (Kulikov et al. 2007; von Paris et al. 2008), which resulted in a less expanded upper atmosphere and exobase levels below the magnetopause. Otherwise early Earth's atmosphere may have been unprotected against solar wind erosion, which should have picked up the planet's initial nitrogen inventory after a few Myr.

Table 8 Exobase altitudes normalized to the solar EUV flux for Venus (96% CO_2 atmosphere) compared to Earth with its present time N_2-rich atmosphere in units of km and Earth radii $r_\oplus$ (after Lammer et al. 2008; Tian et al. 2008)

Composition	96% CO_2/Venus-like	78% N_2 and 22% O_2/Earth-like
EUV/EUV_{Sun}	r_{exo} (km) | [$r_\oplus$]	r_{exo} (km) | [$r_\oplus$]
1	210 | 0.033	500 | 0.078
7	230 | 0.036	15,460 | 2.43
10	250 | 0.04	31,100 | 4.8
20	300 | 0.047	80,800 | 12.7

The present time magnetopause of the Earth is located at about 10 Earth radii

5 Evolution of habitable planets

5.1 Class I habitats: main factors for the evolution of Earth-analog planets

From the discussions in the previous sections we conclude that Class I habitable planets where complex multi-cellular surface life forms as we know on Earth can evolve also need to orbit around the right star (see Fig. 8).

G-type stars and K and F-types with masses close to G stars should fall in this category. In such a case the activity of the host star decreases fast enough so that an evolving atmosphere and life may not be in danger of losing the atmosphere or the planet's water inventory. Furthermore, the large distance of the corresponding HZ of such star systems lessen the efficiency of the non-thermal loss processes. The possibility that various atmosphere compositions and the water inventory can remain stable on such planets over geological time spans exists as long as the environment can keep plate tectonics with all its related consequences active over billions of years.

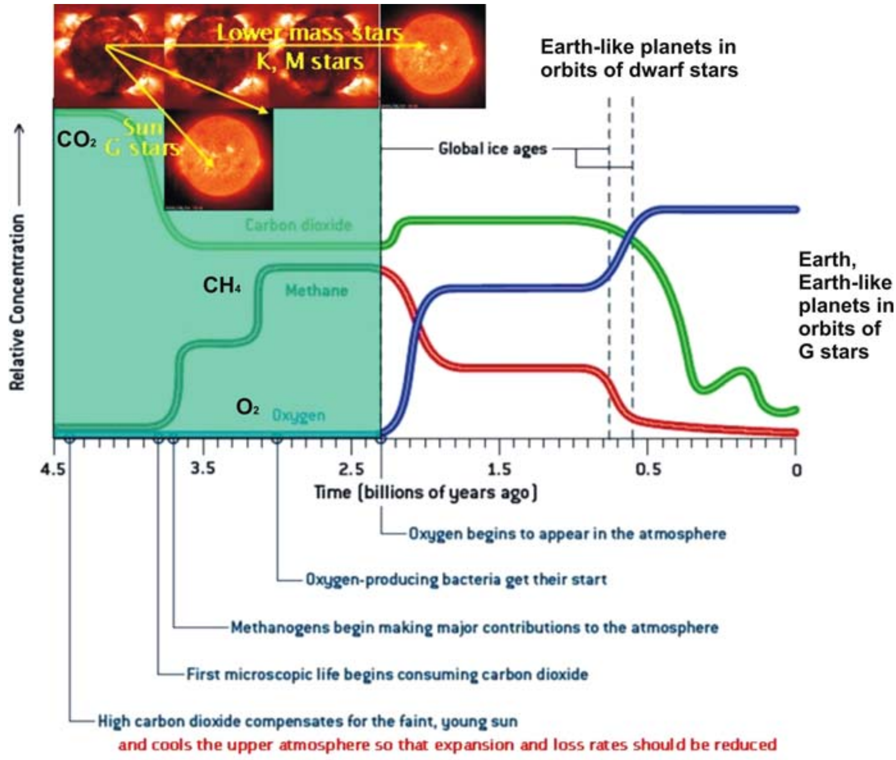


Fig. 8 Illustration of the evolution of Class I habitable planets. These are stellar and geophysical environments where we know from our own experience on Earth that life if originated, can evolve to complex life forms which produce complex interrelated biospheres that may be detected spectroscopically in the future (adapted from Kasting 2004)

5.2 Class II habitats: planetary environments where life may originate but a planet evolves differently from Earth

One can expect that there may be many terrestrial planets orbiting within HZs of low mass M and K-type stars which are located very close to these stars, so that their atmosphere-magnetosphere environments experience extreme stellar radiation and plasma exposures over very long time periods or even during most of their life-time. In such cases thermal and non-thermal atmospheric escape processes could modify the atmospheres and water inventories of the planets in such a way that they may end up after some hundreds of million years as geophysically inactive, dry Venus-like or cold Martian-like planets, although they originated and orbit within the classically defined HZ.

5.2.1 Astrobiological relevance of Class II habitats

The question of whether life could evolve on class II habitats is determined by whether there were habitats available and whether life could have evolved on a water-rich early Venus or early Mars. It is very likely that there are many terrestrial exoplanets which may have obtained a more or less similar or comparable to early Earth volatile inventory during accretion. Then one might expect the possibility of wet-warm water-rich planets where some of them are cooler and others are more Venus-like with close to boiling "moist greenhouse" scenarios (Kasting 1988, 1992).

However, another similarity to present day Venus is that many Class II habitable planets will be tidally locked or slowly rotating planets. As discussed above, slow rotating planets should experience a reduction of the magnetic dynamo as long as the planet is geophysically active and even more so if they are not geophysically active. On the surface of present Venus liquid water does not exist because the temperature is above the critical temperature of about 374°C for pure H_2O and 420°C for salty seawater. Exoplanets with a surface temperature below this critical temperature might host stable liquid water conditions under suitable surface pressure conditions. Venus-like exoplanets with surface temperatures below 150°C are of biological interest (Cockell 1999).

Depending on the spectral type and mass of the host star, water-rich surface conditions may exist for several hundred million years until thermal and non-thermal atmospheric escape processes, together with climate change, modify the planet's environment in a way that the oceans and atmospheres are lost to space. Due to weaker magnetic protection of the atmospheres of such planets, the water and atmosphere can be lost. As discussed in Sec. 4.2 the loss of water affects plate tectonic activity and dynamo generation in a negative way. These scenarios produce less atmosphere protection from the magnetosphere so that only the ionosphere provides protection against atmospheric erosion as outlined in the previous section.

After surface water is lost by various escape processes the formation of carbonates decreases, allowing the increase of CO_2 partial pressure due to volcanic out-gassing so that the planet may evolve to a more Venus-like body. Cockell (1999) investigated the possibilities of Venus-type habitats, past and present. Table 9 summarizes extreme parameters which are of astrobiological importance for Venus-like planets or planets

Table 9 Astrobiologically relevant parameters for CO_2-rich planets compared to biological limits known on Earth (Cockell 1999)

Parameters	Constraints on moist greenhouse planets	Requirements for life on Venus-type exoplanets
Temperature (°C)	Ocean temperatures may be <100°	Temperatures <150° could be taken as an upper limit
Surface pressure (MPa)	No constraint	The temperature is likely to be a limitation before the pressure
Acidity (pH)	pH ≥ 0	pH ≥ 0
UV radiation	UV radiation probably does not prevent life as life could exist within shielded rocks	Absolute upper limit for exposed life is unknown
Atmospheric composition	Same as limits for Earth	Depends on atmospheric composition and possible presence of toxic gases, but a high CO_2 content will not prevent life
Water availability, liquid H_2O on the surface	Water available as hot water bodies	Liquid water must be available. The amount in comparison to other substances depends on the presence of salts and the water activity

which evolved to Venus-like worlds, with a comparison to known biological limits summarized in Table 1 on Earth.

One can also see from Table 9 that several astrobiological aspects emerge which are relevant to the view of early Venus. Unlike present day Venus, the surface temperature range and the water availability on early Venus would have been most likely within the bounds for life (Kasting 1988; Cockell 1999).

Furthermore, there is evidence that life on Earth originated from thermophiles, particularly the greater impactor flux would have provided a physical selection pressure for organisms with a high temperature tolerance (e.g., Sleep et al. 1989). These conditions might favor environments for thermophiles. The possibility of hot water oceans on a moist greenhouse early Venus and high volcanic activity may even have made early Venus a better candidate for the origin of life than early Mars.

Although early Mars may have possessed early and permanent hot water bodies that fit within the region of thermophily with temperatures higher than 50°C, recent data obtained by ESA's Mars Express OMEGA-instrument indicate that there may be no large amounts of carbonates on the Martian surface (Bibring et al. 2005). This result also suggests that a fast and strong early escape triggered most likely by impact erosion of most of the planet's CO_2 atmosphere may have occurred during the early bombardment period (Kulikov et al. 2007).

A recent 3-D MHD model applied by Terada et al. (2009) to a Martian CO_2-rich atmosphere exposed to a 100 times higher solar EUV flux and about 300 times denser solar wind is also in agreement with the impact erosion scenario. Even if one assumes a late onset of a planetary magnetic dynamo, so that no intrinsic magnetic field is taken into account, the loss of the main atmosphere is difficult to explain due to atmospheric loss processes different than impact erosion (Terada et al. 2009). Because of the extreme solar wind interaction with the upper atmosphere and high EUV flux of the young Sun, an ionized obstacle as discussed in Sect. 4.4 at an altitude of $\sim$1,000 km is generated on the dayside of the planet. At this altitude the density of O_2^+ and CO_2^+ and of other heavy ions is very low compared with the O^+ density. Because the solar wind plasma cannot penetrate to altitudes where it can erode the ionized molecules, this results in negligible loss rates of O_2^+ and CO_2^+ ions from early Mars during the active period of the young Sun.

Nevertheless, oxygen ions O^+ can be eroded by the dense early solar wind, resulting in a loss rate capable of eroding not much more than about 8 m of a global Martian ocean. An estimate of the maximum cool ion outflow loss rate is obtained by Terada et al. (2009) under the upper limit assumption that transport of momentum from the early solar wind was efficient enough to accelerate ionospheric ions above the escape velocity through the whole circular ring area around the terminator of the planet. It resulted in an equivalent loss of a global Martian ocean with a depth of $\leq$70 m during the first 100–150 Myr. One can thus see from the study of Terada et al. (2009) that Mars could have lost large quantities of water over its history, but a denser initial CO_2 atmosphere may have been lost only due to some different loss mechanism, most likely due to impact atmospheric erosion (e.g., Walker 1986; Melosh and Vickery 1989; Manning et al. 2006; Pham et al. 2009).

Compared to larger and more massive CO_2-rich planets which may evolve to class II type habitable planets, smaller size planets like Mars cool down faster and water may condense earlier compared to an Earth-size and mass planet. It is expected that the very early Mars should have possessed all the conditions that were also available on early Earth (Westall 2005). These environments may have included deep-water environments below wave basins of standing water (e.g., Cabrol and Grin 2005; Westall et al. 2006) and numerous shallow water and littoral environments around basin edges.

Volcanic and hydrothermal activity would also have been available on early Mars, although not frequent compared to the early Earth (Westall 2005). Figure 9 illustrates

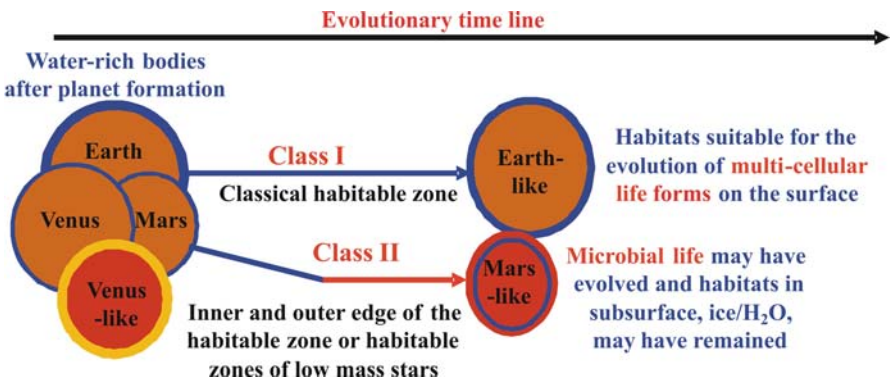


Fig. 9 Illustration of planets which originate within their HZs similar as Class I habitable planets, but due to tidal locking and related weaker magnetic fields, different geodynamic evolution, and strong atmospheric loss processes they may evolve to CO_2-rich Venus-type or Martian-like class II planets where life may have originated but hostile surface conditions prevent the evolution of Earth-like multi-cellular life forms (surface plants, large animals, etc.)

the evolutionary paths of Class II habitats which may originate just like Class I Earth-like planets but due to the closer orbital location of their HZ around dwarf stars they most likely evolve to Class II habitats.

Depending on the spectral type of the star, orbital distance and the related efficiency of atmospheric loss processes, liquid water bodies can be rapidly frozen. The question in such cases is if the liquid water existed for significantly long periods to have been biologically useful, because a frozen surface may not necessarily have inhibited life.

If liquid layers exist below ice layers and these water-reservoirs are in contact with heat sources from the interior of the planet by radioactive, volcanic, or hydrothermal activity, the planet may be considered a Class III habitat, which is discussed in the next section.

5.3 Class III habitats: planetary bodies with subsurface oceans which interact with silicates

Large satellites of gas giants at orbits beyond the ice-line like Jupiter or Saturn are known to include a large amount of water. Indeed, average density being around 1.8, the largest icy moon is composed of almost 45% wt of water. Thus, icy layers are very thick and one can assume reasonably that the thickness is on the order of 600 km for Ganymede, Callisto, and Titan. For Callisto, Galileo data indicate that it is probably not fully differentiated, which means that the icy layer might be thicker, but mixed with silicates. Finally Europa, the smallest of the Galilean satellites, presents a smaller icy layer (100 km), because the amount of water is "only" 10% wt. Understanding the internal structure of the water layer requires a good understanding of the water phase diagram under pressure and temperature as shown in Fig. 10.

Figure 10 shows that many ice polymorphs exist in the moderate pressure range 0–2 GPa, relevant to icy moons. Its other peculiarity, which is of fundamental importance for the icy moons of Jupiter and Saturn, is that the melting curve of the low-pressure ice Ih decreases with pressure, which is related thermodynamically to the fact that ice Ih is less dense than its liquid. The other ices which may coexist with the liquid (III, V, and VI) are denser and present a melting temperature which increases with pressure. From this diagram, it is possible to distinguish four different structures of the thick water layer within icy moons. If the temperature is below 250 K throughout the layer, then it is completely frozen (Case 1 in Fig. 11).

Between 250 and 273 K, it is possible to have a liquid layer located below an icy layer of ice Ih (cases 3–4). If the pressure at the bottom is higher than the melting curve of high pressure ices, then the liquid layer is trapped between the ice Ih layer at the top and a high pressure ice layer at the bottom as shown in case 2 in Fig. 11. Finally, if the temperature is above 273 K, it is then possible to imagine a surface ocean like on Earth. This case, of course, is incompatible with icy moons which present very cold surfaces, but it would be relevant in the case of giant exoplanets with an atmosphere.

Looking at the pressure range encountered within icy moons (Fig. 10), structures 1 and 2 are highly probable for the largest moons, while structure 1 and 3–4 are more probable for Europa (Fig. 11).

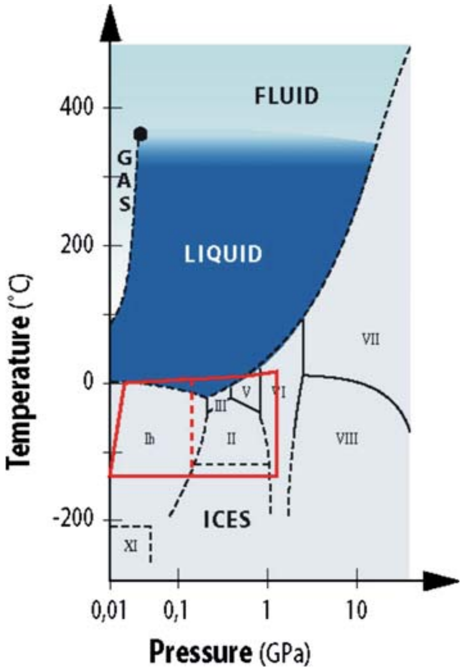


Fig. 10 Phase diagram of the pure water system. Five ice polymorphs can exist in the pressure range relevant to icy moons (regions surrounded by the parallelogram). The *dashed line* symbolizes the highest pressure which can occur in the water layer of Europa

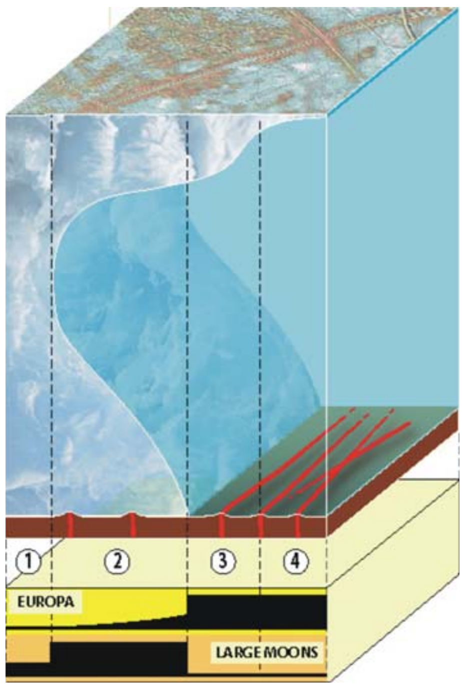


Fig. 11 Possible locations of liquid layers in the icy moons of Jupiter are plotted here as a function of depth: *1* completely frozen, *2* three-layered structures impeding any contact between the liquid layer and the silicate floor, *3* thick upper icy layer (>10 km) and a thick ocean, *4* very thin upper icy layer (3–4 km). Structures 3–4 are the most probable for Europa. The larger moons Ganymede and Callisto are located in the left region (1 or 2) where internal pressures are sufficient to allow for the formation of high-pressure ice-phases. Oceans in Ganymede and Callisto—if they exist—should be enclosed between thick ice layers

Europa is probably unique in that it is the only satellite on which a large ocean can be in contact with the silicate layer. On the other moons, the existence of an ocean implies necessarily the occurrence of a very thick high pressure icy layer at the bottom which impedes the contact of the liquid with silicates. But one must keep in mind that, if the amount of water is large (roughly more than 5% wt of the planet), the liquid layer is still separated from the silicate by a thick high pressure icy layer (i.e., case 2).

Icy and liquid layers are probably not composed of pure ice only. It is assumed that salty materials such as epsomite or natron are trapped within Europa. Many other compounds such as CO_2 (Ganymede and Callisto), or CH_4 (Titan) have been observed on the surface and may issue from the deep interior of the moon. Ammonia is also suspected on Titan. These compounds add a certain amount of complexity to the simple model proposed above, but do not change its conclusions. Their common effects are:

– to reduce the melting temperature of the ice,

and

– to allow for the formation of clathrate and hydrate structures which are trapped within the ices. The first point is actually in favor of larger liquid layers within the satellites.

5.3.1 Astrobiological potential for Class III habitats

The astrobiological potential and possible habitability of Class III habitats rests on the fulfilment of four conditions shown in Fig. 12:

– the presence of liquid water,
– an adequate energy source to sustain the necessary metabolic reactions,

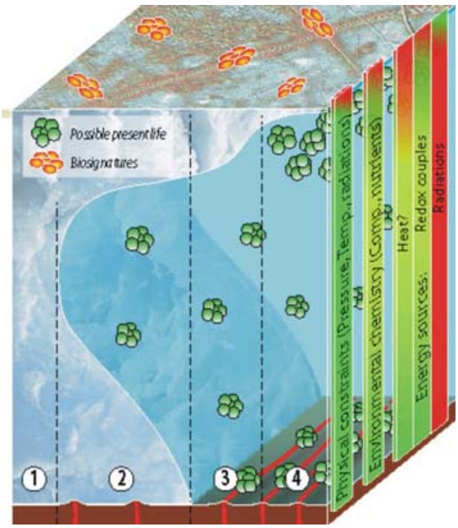


Fig. 12 Present habitability of Europa. Possible locations of present life and biosignatures have been plotted as a function of depth. Habitability depends on physical and chemical constraints which are indicated on the right using color scales (*green* highly favorable, *red* hostile). *Numbers* refer to possible interior structures described in Fig. 11

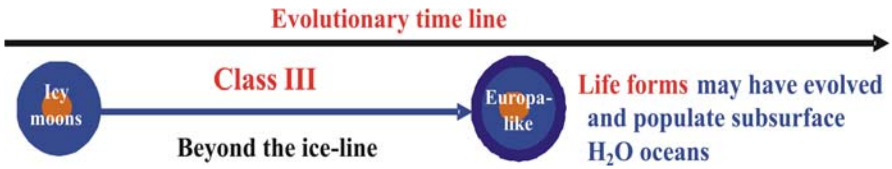


Fig. 13 Illustration of icy satellites of gas giants (Class III habitats) which have liquid water habitats or oceans beyond an ice-layer which interacts with the silicate core or sea floor. Life may have originated within such subsurface habitats but surface conditions are hostile

and

– a source of chemical elements (C, N, H, O, P, S, etc.), which can be used as nutrients for the synthesis of bio-molecules, and
– relevant pressure and temperature conditions.

The fulfilment of the first condition is directly related to the putative existence of a subsurface ocean. The existence of such an ocean within Europa has been inferred from the interpretation of the induced magnetic field measured by the Galileo mission and the interpretation of geological features. If it exists, its depth is unknown. It could be very close to the surface (1–2 km) or much deeper (20 km or more). The interface between the icy crust and the ocean is expected to be relatively flat: since it corresponds to a thermodynamical equilibrium between a solid ice and its liquid counterpart. Temperature and pressure along the interface must then be almost constant, which impede significant depth variations. Furthermore, the flat topography does not suggest the presence of large roots of ice below high mountains as seen on Earth.

According to current models, it may be the only case in the Solar System in which liquid water is globally in contact with a silicate core (see also Fig. 13). Such conditions are favorable for interactions between the subsurface ocean and silicates, particularly if a volcanic and consequently hydrothermal activity exist.

These unique conditions on Europa allow water-rock interactions, especially if any volcanism exists. The determination of the topography and/or mass anomalies at the silicate core/liquid interface would provide hints on whether volcanism exists or has existed.

This provides a variety of chemicals and energy that could play a role in sustaining putative life forms at the ocean floor like those discovered 30 years ago at Earth deep-sea hydrothermal vents.

However, this task may prove difficult, mainly because the presence of the icy crust above Europa's liquid layer decouples the surface topography from the ocean floor, but it is an objective which will possibly be fulfilled by future space missions to the Jupiter system.

5.4 Class IV habitats: subsurface-ice water worlds and "Ocean planets"

We define as Class IV habitats planetary satellites or planets which have subsurface water oceans, water reservoirs or oceans on the surface which do not interact with a typical silicate bearing sea floor like the silicate core on Europa (e.g., case 2 in Fig. 11). Subsurface oceans may also exist between different ice-layers at Ganymede

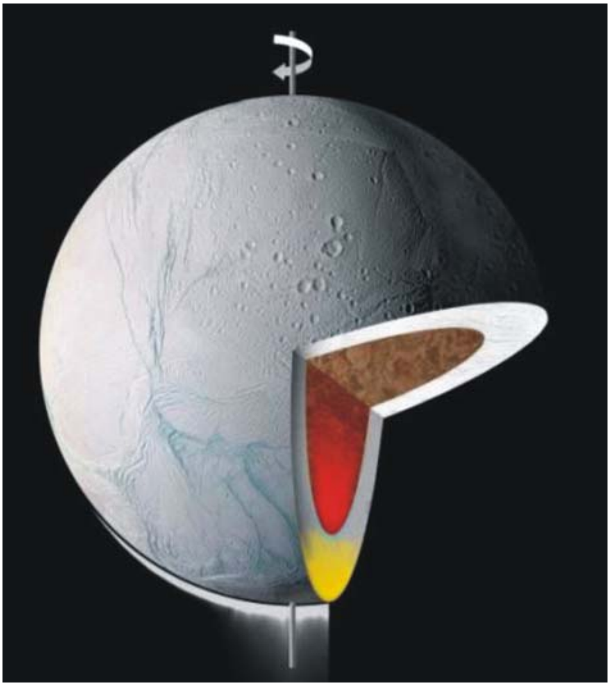


Fig. 14 Possible internal structure for Enceladus (NASA/JPL)

and Callisto. In addition to the Galilean moons, NASA's Cassini spacecraft has shown that Saturn's moon Enceladus possesses active plumes, the sources of which may be pockets of liquid water near the surface, an intriguing possibility for such a small and distant satellite (Fig. 14). The jets emanating from the South Pole of Enceladus are almost certainly the most accessible samples from a liquid water environment in the outer Solar System.

5.4.1 Enceladus as an example of a Class IV habitat

The Ion Neutral Mass Spectrometer (INMS) instrument on board Cassini has found non-condensible volatile species (e.g., N_2, CO_2, CH_4) in jet-like plumes over Enceladus' geologically active South Polar Terrain (see Fig. 14 and Waite et al. 2007). As illustrated in Figs. 14 and 15 the most likely source of Enceladus' jets is a pressurized subsurface liquid water reservoir. If N_2 is present it may reflect thermal decomposition of ammonia associated with the subsurface liquid reservoir and may imply that the water has some exchange with hot rocks—providing a source of heat as well as mineral surfaces for catalyzing reactions. If this scenario proves correct, then all the ingredients are present on Enceladus for the origin of life by chemoautotrophic pathways—one possible model for the origin of life on Earth in deep sea vents. In this model, life on Earth began in deep sea hot springs where chemical energy was available from a mix of H_2, S and Fe compounds.

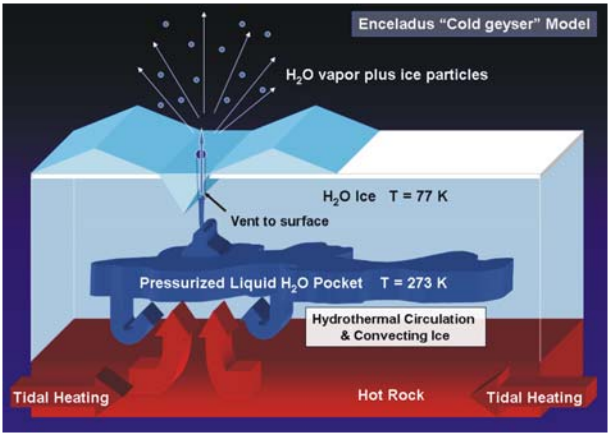


Fig. 15 Model of Enceladus plumes (NASA/JPL/SSI)

The fact that the branches of the tree of life that are closest to the common ancestor are thermophilic has been used to argue that this reflects a thermophilic origin of life—although other explanations are possible. In the material emanating from Enceladus we can search for the specific molecules associated with such systems, including H_2 H_2S, FeS, etc. and the expected organic byproducts of the ecosystems that would inhabit such systems—biomarkers. Interestingly, CH_4 is the key expected biological by-product, although there are many non-biological ways CH_4 can be produced.

The low molecular weight organics detected by Cassini may be just one part of a suite of organics present in the plume and on the surface. Studies of the nature of these organics can tell us if they are produced biologically or non-biologically.

The molecular species likely to be produced by such a prebiotic or biotic chemistry—such as amino-acids, lipid compounds and sugars—could be detected in the plumes of Enceladus using in situ techniques. It is also crucial to confirm the presence of liquid water reservoirs, both by remote sensing measurements and measurements taken on the surface. A key question is whether Enceladus formed from primordial material similar to that of Titan.

In contrast to the Jovian system, characterized by a negative gradient in the satellite uncompressed density with distance from Jupiter, the Saturnian system exhibits large variations in mean density and mass of the satellites. So we should expect great variability also in the satellites' bulk chemical composition. Some species, as for instance CO_2, should be primordial, whereas others, like N_2, would require a liquid water reservoir in the interior at elevated temperatures, enabling aqueous, catalytic chemical reactions (Matson et al. 2007). Taken together, the evidence strongly suggests that Enceladus' interior is at least partially differentiated into a silicate-metal core overlain by a water-ice liquid shell (Schubert et al. 2007).

However, the energy source required to initiate and preserve activity at the level comparable to that observed today over geologic periods and the concentration of geologic and thermal activity towards the south-polar region are still not well understood. On Enceladus, altimetric and gravimetric profiles performed during close flybys

(<100 km) over the South Pole can be used to retrieve the internal structure below the active region (depth of liquid reservoirs, lithosphere thickness, existence of thermal plumes in the ice mantle, in the rocky core) (e.g., Schubert et al. 2007). These data will provide fundamental constraints on the origin of the South Pole hotspot on Enceladus. However, most of the important astrobiological aspects will not be answered by Cassini-Huygens and require a dedicated new mission to Titan and Enceladus.

5.4.2 Titan's subsurface ocean and surface lakes

Another planetary body which has even more similarities with the Earth (Coustenis and Taylor 2008), but can be considered as a possible Class IV habitat, is Saturn's largest satellite Titan, which is unique in the Solar System due to its extensive nitrogen atmosphere, four times denser at the surface than on our own atmosphere, and host to a rich organic chemistry and some CH_4. Titan's atmosphere is not in chemical equilibrium. In it, a chemical factory initiates the formation of complex positive and negative ions in the high thermosphere as a consequence of magnetospheric–ionospheric–atmospheric interactions. The second most abundant constituent, methane, is dissociated irreversibly to produce hydrocarbons and nitriles, by combination with nitrogen. Recent Cassini-Huygens discoveries have revolutionized our understanding of the Titan system and its potential for harboring the "ingredients" necessary for life (Coustenis and Taylor 2008). These discoveries reveal that Titan is rich in organics, most probably contains a vast subsurface ocean, and has sufficient energy sources to drive chemical evolution.

As recently discovered by the Cassini INMS, in the high atmosphere, heavy ions are formed (Waite et al. 2007). This energetic chemistry produces large molecules like benzene, which begin to condense out at ~950 km, are detected as neutral hydrocarbons and nitriles in the stratosphere and initiate the process of haze formation. As the haze particles fall through the atmosphere and grow, they become detectable with imaging systems such as the Cassini/ISS at ~500 km and are ubiquitous throughout the stratosphere. They are strong absorbers of solar UV and visible radiation and play a fundamental role in heating Titan's stratosphere and driving wind systems in the middle atmosphere, much as ozone does in the Earth's middle atmosphere. Eventually, these complex organic molecules are deposited on Titan's surface in large quantities, where data from Cassini's instruments hint at their existence. Hence the upper thermosphere is linked intimately with the surface and the intervening atmosphere.

Surprisingly, the Gas Chromatograph and Mass Spectrometer (GCMS) on board Huygens did not detect a large variety of organic compounds in the low atmosphere. The mass spectra collected during the descent show that the medium-altitude and low stratosphere and the troposphere are poor in volatile organic species, with the exception of methane (Niemann et al. 2005). Condensation of these species on aerosol particles is a probable explanation for these atmospheric characteristics. These particles, for which no direct data on their chemical composition were previously available, were analyzed by the ACP instrument. ACP results show that the aerosol particles are made of refractory organics which release HCN and NH_3 during pyrolysis. This supports the tholin hypothesis ("tholins" are solid products in the laboratory mimicking complex refractory organics: Nguyen et al. 2007; and references therein). From these new

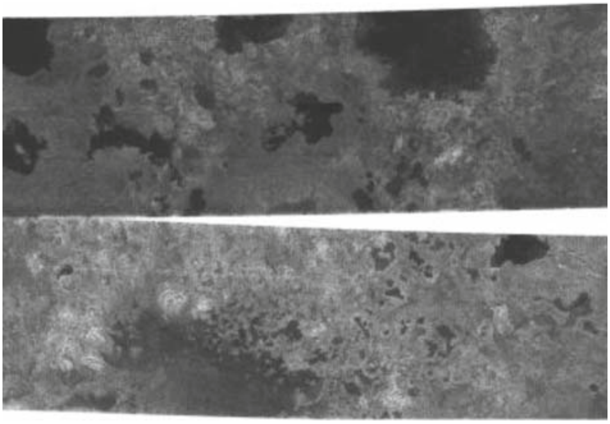


Fig. 16 Lakes discovered on Titan by the Cassini–Huygens mission (NASA)

in situ measurements it seems very likely that the aerosol particles are made of a refractory organic nucleus, covered with condensed volatile compounds (Israël et al. 2005). However, the nature and abundances of the condensates have not been measured. Even more importantly for astrobiology, neither the elemental composition nor the molecular structure of the refractory part of the aerosols has been determined.

After several close Cassini flybys, in January 2005, the Huygens Probe became the first human artefact to descend through Titan's atmosphere, reach the surface and return several hours of data from an exotic landscape cut by channels and apparently soaked in the near-surface with ethane and methane. Several years of flybys by the Cassini orbiter have led to radar and near-IR maps that show dunes made of icy and organic fine grains extending for thousands of kilometers. Titan was once thought to conceal a global ocean of methane and ethane beneath its smoggy atmosphere. However, while the landforms seen by Cassini and Huygens show ample evidence of past modification by the action of flowing liquids, actual bodies of present liquid have proven elusive through more than two years of investigation, until the July 22, 2006 flyby, when Cassini's RADAR instrument finally unveiled what appears to be a land of lakes in Titan's northern polar regions (see Fig. 16). ISS images also show a kidney-shaped dark feature about 200 km in length, named Ontario Lacus, that is outside the area of radar coverage and has recently been confirmed by the VIMS instrument not only to be a lake, but also to be composed of ethane liquid (Brown et al. 2008). In the absence of a massive surface ocean but with analogs to all other terrestrial hydrological phenomena present, Titan's methane cycle is very exotic.

Indeed, on Titan, methane can exist as a gas, liquid and solid. Playing a role similar to that of water on the Earth, methane is cycled between the atmosphere and the surface.

Cloud systems, the size of terrestrial hurricanes (1,000 km large), appear occasionally, while smaller systems are there on a daily basis. Titan's atmospheric methane may be supplemented by high latitude lakes and seas of methane and ethane, which over time cycle methane back into the atmosphere where it rains out, creating fluvial erosion over a wide range of latitudes. With the current picture of Titan's organic chemistry, the chemical evolution of the main atmospheric constituents—dinitrogen and methane—produces mainly ethane, which accumulates on the surface or the near subsurface, eventually dissolved to form methane–ethane lakes and seas, and complex refractory organics which accumulate on the surface, together with condensed volatile organic compounds such as HCN and benzene.

In spite of the low temperature, Titan is not a congealed Earth: the chemical system is not frozen. Titan is an evolving planetary body and so is its chemistry. Once deposited on Titan's surface, the aerosols and their complex organic content may follow a chemical evolution of astrobiological interest. Laboratory experiments show that, once in contact with liquid water, Titan tholins can release many compounds of biological interest, such as amino acids (Khare et al. 1986). Such processes could be particularly favorable in zones of Titan's surface where cryovolcanism is occurring. The N_2–CH_4 by-products in Titan's atmosphere eventually end up as sediments on the surface, where they accumulate presently at a rate of roughly 0.5 km in 4.5 Gyr.

Thus one can envision the possible presence of such compounds on Titan's surface or near subsurface. Long-term chemical evolution is impossible to study in the laboratory: in situ measurement of Titan's surface thus offers a unique opportunity to study some of the many processes which could have been involved in prebiotic chemistry, including isotopic and enantiomeric fractionation (Nguyen et al. 2007).

Even with the detection of the large lakes in the North, no viable source was detected by Cassini to re-supply methane. Cryovolcanic outgassing has been hypothesized, yet over what timescales and through which internal processes are unknown. Cassini-Huygens also found that the balance of geologic processes—impacts, tectonics, fluvial, Aeolian—is somewhat similar to the Earth's, more so than for Venus or Mars. Titan may well be the best analog to an active terrestrial planet in the sense of our home planet, albeit with different working materials.

Although Titan is much colder than the Earth, it does present many similarities with our planet (see Fig. 17). Titan's atmosphere is made of the same main constituent, dinitrogen. It also has a similar structure from the troposphere to the ionosphere, and a surface pressure of 1.5 bar—the only case of an extraterrestrial planetary atmospheric pressure close to that of Earth. As noted earlier, many analogies can also be made between the role of methane on Titan and that of water on the Earth.

Methane on Titan seems to play the role of water on the Earth, with a complex cycle that has yet to be fully understood. Analogies can also be made between the current organic chemistry on Titan and the prebiotic chemistry which was active on the primitive Earth. In spite of the absence of permanent bodies of liquid water on Titan's surface, the chemistry is similar.

Moreover, Titan is the only planetary object, other than the Earth with long-standing bodies of liquid on its surface (although direct observational evidence of the longevity of Titan's surface liquids remains to be obtained). Several of the organic processes which are occurring today on Titan form some of the organic compounds which are considered as key molecules in terrestrial prebiotic chemistry, such as hydrogen cyanide (HCN), cyanoacetylene (HC_3N) and cyanogen (C_2N_2). In fact, with several percent of methane in dinitrogen, the atmosphere of Titan is one of the most favorable atmospheres for prebiotic synthesis.

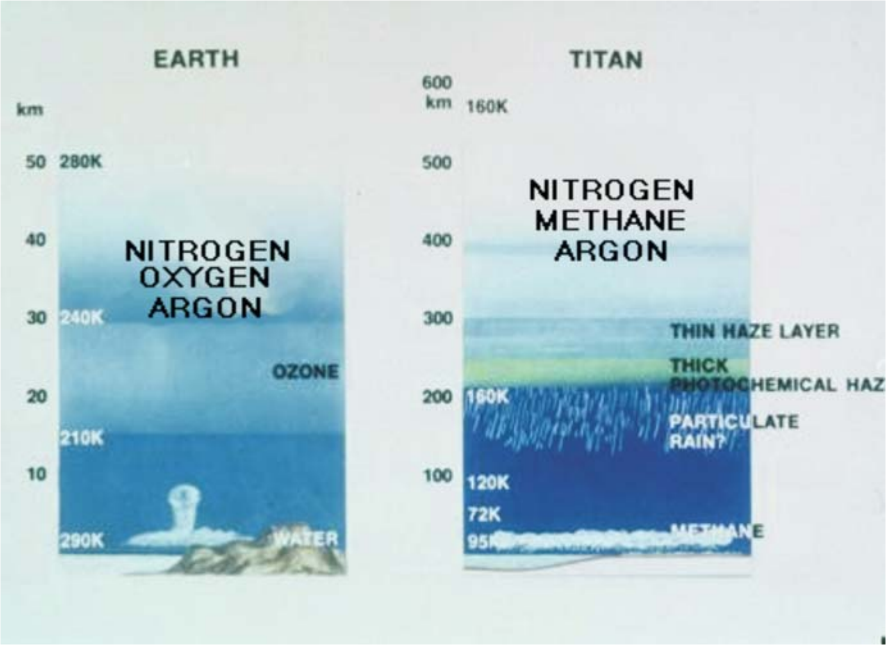


Fig. 17 Atmospheric structure of Titan and the Earth

The degree of complexity that can be reached from organic chemistry in the absence of permanent liquid water bodies on Titan's surface is still unknown, but it could be quite high. All ingredients that are supposed to be necessary for life to appear and develop—liquid water, organic matter and energy—seem to be present on Titan. Indeed, interior structure models suggest that Titan, like Europa, Ganymede and Callisto, has maintained internal liquid water reservoirs (probably mixed with some ammonia and more speculatively sulfur) (see Fig. 18). At the beginning of Titan's history, this hypothetical subsurface ocean may have been in direct contact with the atmosphere and with the internal bedrock, offering interesting analogies with the primitive Earth, and the potential implication of hydrothermal vents with terrestrial-like prebiotic chemistry.

Recent models of Titan's interior including thermal evolution simulations predict that the satellite may have an ice crust between 50- and 150-km thick, lying atop a liquid water ocean a couple of hundred kilometers deep, with some amount (a few to 30%, most likely ∼10%) of ammonia dissolved in it, acting as an antifreeze. Beneath lies a layer of high-pressure ice (Fig. 18). Indeed, Cassini's measurement of a small but significant asynchronicity in Titan's rotation is most straightforwardly explained by the separation of the crust from the deeper interior by a liquid layer (Lorenz et al. 2008). Definite detection of this ocean of water and ammonia under an icy layer can be provided by the Radio Science Subsystem on board Cassini, by measuring the principal components of Titan's and Enceladus gravitational potential (Rappaport et al. 1997, 2007). This will provide important constraints on the satellites' internal differ-

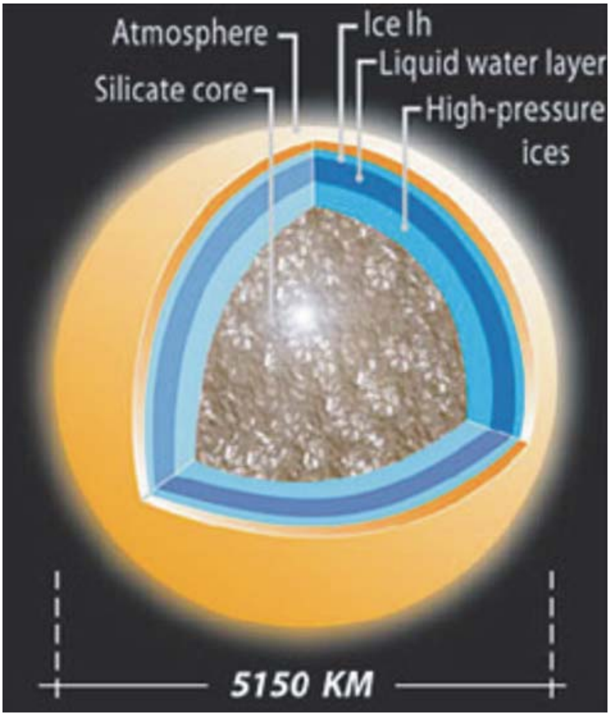


Fig. 18 Illustration of Titan's internal structure with a liquid ocean between two subsurface-ice layers (from Tobie et al. 2005)

entiation. However, the size of their cores and the thicknesses of their ice mantles will remain uncertain (Sohl et al. 2003). Determining the present-day structure of the outer H_2O ice mantle and of the innermost core is crucial to determining how a satellite's interior is differentiated over time and its effects on the evolution of the surface and atmosphere.

Although the chemical reactions that lead to life on Earth take place in liquid water, the reactions themselves are almost entirely between organics. The study of organic chemistry is therefore an important, and arguably richer, adjunct to the pursuit of liquid water in the Solar System. Titan's organic inventory is impressive, and carbon-bearing compounds are widespread across the surface in the form of lakes, seas, dunes and probably sedimentary deltas at the mouths of channels.

The Cassini-Huygens mission has significantly enhanced our understanding of Titan as the largest abiotic organic factory in the Solar System. The abundance of methane and its organic products in the atmosphere, seas and dunes exceeds the carbon inventory in the Earth's ocean, biosphere and fossil fuel reservoirs by more than an order of magnitude (Lorenz et al. 2008). The Cassini/INMS showed that the process of aerosol formation appears to start more than 1,000 km above the surface through a complex interplay of ion and neutral chemistry initiated by energetic photon and particle bombardment of the atmosphere (Waite et al. 2007), and includes polymers of high molecular weight—up to and certainly beyond C7 hydrocarbons, which the

Cassini mass spectrometer is unable to measure. Measurements throughout the atmosphere, both remotely and in situ, have indicated the presence of numerous hydrocarbon and nitrile gases, as well as a complex layering of organic aerosols that persists all the way down to the surface of the moon. (Coustenis et al. 2007; Tomasko et al. 2005). Radar observations suggest that the ultimate fate of this aerosol "rain" is the generation of expansive organic dunes that produce an equatorial belt around the surface.

A significant geophysical difference evident when one compares Titan and Europa is that on Titan the liquid water is not presently in contact with a silicate core (on Titan the core is isolated from the ocean by a layer of a high-pressure ice phase, although in the past there could have been intimate mixing of silicates with liquid). The surface of Titan appears (like the surface of Mars or Europa) an unlikely location for extant life, at least for terrestrial-type life. However, it has been noted (Fortes 2000) that Titan's internal water ocean might support terrestrial-type life that had been introduced there previously or formed when liquid water was in contact with silicates early in Titan's history. McKay and Smith (2005) have noted that there are photochemically derived sources of free energy on Titan's surface which could support life, which would have to be an exotic type of life using liquid hydrocarbons as solvents. In a similar vein, Stoker et al. (1990) observed that terrestrial bacteria can in fact satisfy their energy and carbon needs by 'eating' tholin. In this sense, a methane-rich atmosphere may act as a 'poor-planet's photosynthesis', providing a means to capture the free energy from ultraviolet light and make it available for metabolic reactions.

It is clear that Titan's organic chemistry and sub-surface ocean (among other) remain to be investigated. In particular, joint measurements of large-scale and mesoscale topography and gravitational field anomalies on Titan from an orbiter and from an aerial platform would impose important constraints on the thickness of the lithosphere, the presence of mass anomalies at depth and any lateral variation of the ice mantle thickness. It is astrobiologically essential to confirm the presence of such an internal ocean, although the water layer may not be in contact with the silicate core like Europa.

Consequently, it cannot be excluded that life may have emerged on or in Titan. In spite of the extreme conditions in this environment, life may have been able to adapt and to persist. Even the possible current conditions (pH, temperature, pressure, salt concentrations) are not incompatible with life, as we know it on Earth (Fortes 2000). However, the detection of a potential biological activity in the putative liquid mantle seems very challenging.

5.4.3 "Ocean planets"

Besides habitats on planets and satellites in the Solar System which can act as a mirror for extra-solar systems, there may be planets in the Universe which are completely different to those existing in our Solar System. All of the discovered extra-solar planetary systems so far do not resemble our own Solar System. Because it seems reasonable that some form of migration due to interactions between planets and the protoplanetary disk plays a role in young systems (e.g., Lin et al. 1996; Ward 1997; Trilling et al. 1998), Léger et al. (2004) proposed that slightly massive Uranus or Neptune-like planets may have formed beyond the ice-line of a protoplanetary disk and migrated inward to distances where liquid H_2O can be present at their surface. If such planets exist they are certainly interesting bodies for planetology and astrobiology. Léger et al. (2004) suggested that organic molecules and the elements necessary for the origin of life like P, S, Fe, Mg, Na, K, etc. could be brought to the surface by micrometeorites or found in the ocean as dissolved species.

For instance, if biomarkers are discovered spectroscopically in the atmosphere of an "Ocean planet", it would indicate that life on such habitats could have evolved in the absence of black smokers which are not expected on these planets because liquid water and silicates are separated by thousands of kilometers of ice. In stellar systems where planetesimals built in the colder regions of the protoplanetary disk contain a significant fraction of water ice, planets would mimic the icy satellites of the giant planets. Uranus and Neptune can themselves be considered as "ice giants" because the interior of these bodies and their density is very similar to that of compressed water ice (e.g., Podolak et al. 2000).

On the other hand ice giants like Uranus and Neptune also contain about 1–4 Earth masses of hydrogen and helium in the form of an outer gaseous envelope. "Ocean planets" as suggested by Léger et al. (2004) contain much less hydrogen. Planets with masses in the range $1 < M/M_\oplus < 8$ do not accrete large amounts of hydrogen (Wuchterl et al. 2000). "Ocean planets" are supposed to form beyond the iceline, e.g., $\sim$5–10 AU and migrate inward to orbital distance $\leq$1 AU on timescales which are of the order of $\leq$1 Myr (e.g., Lin et al. 1996).

These planets are mostly made of refractory material like metals, silicates and ices. The composition of the protosolar nebula contains about 50% metals and silicates, and 50% water ice. On an "Ocean planet", an ocean is present with a surface temperature which is in between the triple point and critical temperature of $\sim$273–647 K.

The water-silicate interface region corresponds to the temperature of the melting point of ice at the local pressure. For a $6M_\oplus$ "Ocean-Planet, this pressure is $\sim$250 GPa as shown in Table 10 and the corresponding melting point of high pressure ice VII is 1,150 K, as estimated from the Simon fusion equation in Mishima and Endo (1978). Léger et al. (2004) argued that the temperature at this interface is higher than the ice melting temperature and most of the water shell is liquid. In such a case high-pressure/high-density ice would lead to the sinking of and the fast build-up of an ice

Table 10 Calculated internal structure of a 6 $M_\oplus$ Ocean planet

Super Earth	$r[r_\oplus]$	ρ (gcm^{-3})	P (GPa)	$g[g_\oplus]$
Center	0.0	19.5	1,580	0.0
Iron–silicate interface	0.69	15.6	735	2.1
Silicate–ice	1.24	6.2	250	1.96
Surface under the ocean	2.0	1.5	1.0	1.54

The planet consists from the center to the outside of 1 $M_\oplus$ metals, 2 $M_\oplus$ silicates, and 3 $M_\oplus$ ice. The density at the center, different interfaces and top is: 19.5, 15.6–8.2, 6.2–3.9 and 1.54 g cm^{-3}, the gravity is 0, 2.1, 1.96, and 1.54 $g_\oplus$, and the pressure in units of GPa is about 1,580, 735, 250, and 1. The upper layer of the planet is a $\sim$100 km thick ocean layer. The mean planetary density is 4.34 g cm^{-3} (after Léger et al. 2004)

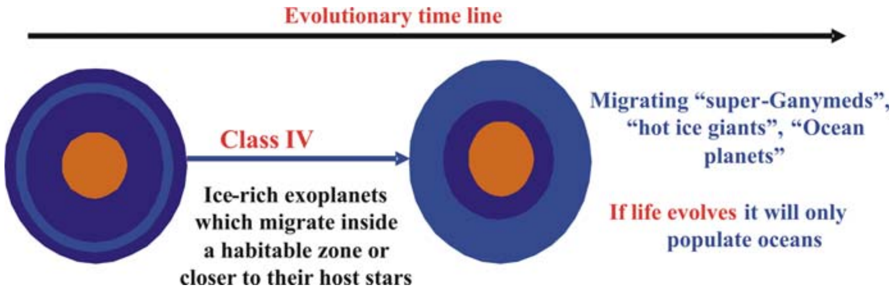


Fig. 19 Illustration of "Ocean planets" which are much bigger than Ganymede, Callisto, Titan and Enceladus but can be also considered as class IV habitats

shell. Furthermore, the temperature at the water-silicate interface is much lower than the melting temperature of ice VII and the internal heat is transferred by subsolidus convection in the ice layer.

A comparison between models built for the internal dynamics of icy satellites of the Solar System gas giants indicate that subsolidus convection in ice is insufficient to remove the heat produced by the decay of radiogenic elements which are contained in the silicate shell (e.g., Grasset et al. 2000). It is expected that the only possibility for transferring heat is the melting of ice and migration of the melt towards the upper ocean. Léger et al. (2004) describe the amount of melt produced by internal heating as being in the order of a 1 km thick layer per Myr at the interface, which is small compared to the size of the ice layer. As a consequence, the temperature profile in the ice mantle of an "Ocean planet" follows the solid–liquid transition curve from 1,150 K at the silicate interface to the temperature at the bottom of the ocean.

Although the behavior of silicates in the pressure domain expected in "Ocean planets" is unknown, one may assume that the silicates are composed of perovskite and magnesowustite and that the fossil heat plus that produced by the decay of long-lived radiogenic elements in the silicate shell, are transferred by subsolidus convection (e.g., Schubert et al. 2000; Léger et al. 2004). In the metal-silicate core, the temperature decreases from a high value in the center like $\sim$10,000 K for a $6M_{\oplus}$ Ocean planet to the ice fusion temperature at the silicate-ice boundary of $\sim$1,150 K. In the ice shell, it follows the water liquid–solid transition to the ocean bottom temperature.

The possibility of such water-rich planets existing raises important astrobiological questions (Fig. 19):

– Can life originate on such bodies, in the absence of continent and ocean–silicate interfaces?
– What would be the nature of an out-gassed atmosphere and the geochemical cycles?
– What is the orbit distance where "Ocean planets" can survive?

In the case of very small orbital distances the liquid/gas interface can disappear, and the hot water envelope is made of a supercritical fluid. The water reservoir of these planets seems to be weakly affected by gravitational escape, provided that they are located beyond some minimum distance, e.g., 0.05 AU for a 5–6 Earth-mass planet around a Sun-like star (Selsis et al. 2007).

6 Origin of life as we know it

6.1 What are minimum requirements for the origin and evolution of life?

Before we start discussing any details on the necessary conditions for life, we need to define life as we know it. Life as we know it has been described as a (thermodynamically) open system (Prigogine et al. 1972), which makes use of gradients in its surroundings to create imperfect copies of itself. A minimal form of life has to fulfill a number of functions, specified by its definition.

As any open thermodynamic system needs a (permeable) boundary. Minimal life needs to separate internal processes from the outside world. This boundary is achieved by a membrane in current (known) biology. An internal machinery to convert an external gradient into energy is necessary to achieve replication. In known terrestrial biological systems this is achieved by proteins, lipids and nucleic acids. To compete well in evolution, the minimal organism also needs a way of passing on information to subsequent generations. In terrestrial biology the genetic material in the form of nucleic acids is responsible for templating. It has been suggested that in order to reach a sufficient degree of complexity, it is necessary to have interdependent but separate structures to fulfill all those functions (Ruiz-Mirazo et al. 2004). Figure 20 shows the minimum requirements for a hypothetical living organism.

The necessary components for a successful origin life and evolution, we need the equivalents of:

– a membrane,
– proteins, lipids and other macromolecular structures,
– an information carrier,

and

– a gradient that can be utilized as an energy source.

And all these need to be interacting to form a replicating system.

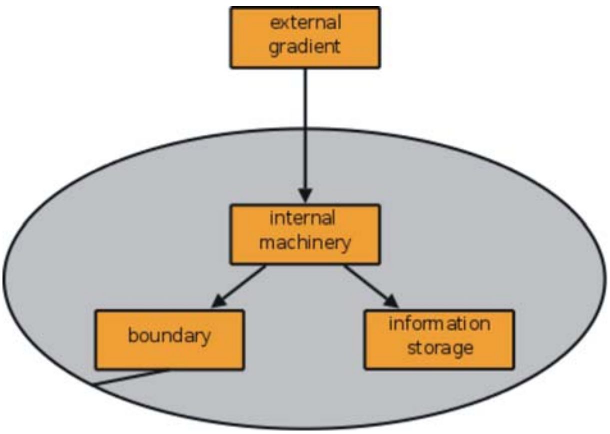


Fig. 20 Schematic illustration of a hypothetic minimal organism and its main functions

Organic molecules and macromolecules, including prebiotic compounds have been delivered to the early Earth, see Sect. 2 via comets, meteorites and interplanetary dust particles during the early formation of the Solar System. As was pointed out in Sect. 2, the inner Solar System was far too hot when the Earth and its terrestrial neighbor planets formed, to allow any of the carbon bearing species to survive the accretion process. Organic material had to be delivered to or be formed on Earth after it had cooled down sufficiently. There are many examples of prebiotic reactions, which form biologically important molecules; best known among these is probably the famous Miller experiment. (Miller 1953; Miller and Urey 1959). It would seem that in all these reactions the exact composition of the starting materials as well as the exact nature of the energy source is of little consequence. The resulting mixture of organic molecules from Miller-type experiments and ice irradiations with various energy sources is quite similar. (Bada and Lazcano 2003; Kiyakawa et al. 2002; Briggs et al. 1992; Holtom et al. 2005). The only exceptions to this rule are experiments that use carbon dioxide CO_2 as the carbon-bearing starting molecule. In this case the main molecule formed is CO_2 (Bredehöft and Meierhenrich 2008). This does not come as a huge surprise, considering that the thermodynamically most stable combination of two equivalents of oxygen and one equivalent of carbon is carbon dioxide, which was most likely a dominant part of the early atmosphere of terrestrial planets (see Sect. 5.1).

In “our” kind of Earth-biology, however, there are four different chemical groups that are essential to life as we know it. While it is tempting to consider them as the only possible ingredients of life, for lack of comparison we do not know whether there are other chemical foundations conceivable. The functions as defined above, their chemical footing on Earth and possible alternatives will be discussed in the following paragraphs.

6.1.1 The lipids of the membrane

On Earth the first group of essential chemicals are the lipids or fatty acids, forming the membranes of all cells. It has been shown, that these structures, or precursors of them can form spontaneously. (Luisi and Varela 1989; Luisi et al. 1999). The driving force for the formation of these proto-cells is simply the increase in entropy when hydrophobic molecules stick together in aqueous surroundings, thus freeing a lot of water molecules that would otherwise be oriented on the surface of the (larger) hydrophobic molecules. Since water is the medium on Earth which facilitated the origin of life we can assume that the formation of micelles and vesicles acted as the outside boundary for the first proto-cells. Compartimentization may have occurred in this way from diverse starting materials (e.g., carboxylic acids, PAHs, lipids, etc.).

6.1.2 The amino acids of proteins

The biochemical architecture and metabolic pathways of all life forms on Earth require proteins, which are composed of amino acids. Perhaps the most striking feature of the biologically used amino acids is their homochirality. Amino acids are chiral, meaning they come in two varieties: left-handed and right-handed called L-enantiomer and D-enantiomer. Life, however, predominantly uses the L-enantiomer for building proteins (D-amino acids are very rarely used in cell receptors and in some enzymes). The origin of homochirality is currently unknown. It has been argued that the origin of homochirality preceded the origin of life.

There is a lot less literature regarding the question of whether other molecules than amino acids could be the building blocks of life‘s polymers. The fact that amino acids can be found in great variety in meteorites (so far at least in our Solar System over 90 individual amino acids have been identified (Bredehöft and Meierhenrich 2008), together with their chemical properties (good solubility and polymerization in water, low vapor pressure, etc.) certainly makes them a good choice, but we do not know whether other molecules creating different polymers might not be an equally effective. There is, however, a certain consensus that the collection of those 20 amino acids found in all life forms is not entirely by chance, but rather that the best option for each chemical function is chosen (Weber and Miller 1981).

6.1.3 The sugars and nucleobases of the genetic material

The last set of chemical components vital to life are the sugars that make up the backbone of Deoxyribonucleic acid (DNA) and the nucleobases that are attached to that backbone, encoding the actual information. DNA is used in biological information storage. In DNA the backbone consists of a chain of alternating molecules of the 5-carbon sugar deoxyribose and phosphoric acid. Individual pairs of one sugar molecule and one molecule of phosphate are called a nucleotide. There are four different nucleotides in DNA, differing in which of the nucleobases adenine, guanine, cytosine or thymine is attached to the sugar. The actual information in DNA is encoded by the sequence those nucleotides form.

It is pretty certain that DNA was not the original primordial form of the molecule that stored genetic information. It may have been preceded by RNA (Gilbert 1986) a molecule made of another sugar, the 5-carbon sugar ribose. The chemical difference between these two sugar molecules might seem only slight (only one hydroxyl-group more in ribose) but the consequences are considerable. RNA is capable of catalyzing its own copy (DNA is not) (Lincoln and Joyce 2009). RNA can also perform other valuable biological functions, namely making proteins. The one advantage that DNA has over RNA is that information encoded in DNA is more stable and better shielded than in RNA. For complex life mutations are less likely to be deleterious because they have more spare cells. In very early complex life forms mutations, alterations of the genetic sequence, created new traits in offspring. RNA also uses one different nucleobase, it utilizes uracil instead of thymine. (Uracil is more easily created than thymine, because it can be formed from cytosine by losing an amino group, but that also means, that the spontaneous alteration of cytosine would alter the genetic code to spell something different. When cytosine mutates in DNA, the specific codon can be sensed as damaged and be repaired. So using thymine is another precaution against unwanted mutation).

RNA may have played a crucial role in early biological evolution and might even have been preceded by an even earlier form of genetic molecule, the so called Peptide Nucleic Acid PNA (Nielsen 1993) which is not based on sugars and phosphoric acid, but on diamino acids. The auto-catalytic properties of PNA have not been explored as extensively as for RNA. One advantage that PNA has over RNA is that its monomers, the diamino acids, can be found in meteorites (Meierhenrich et al. 2004) while neither ribose nor deoxyribose, or for that matter any other sugars have ever been found in prebiotic sources. This is most likely explained by the relatively unstable chemical nature of sugars and related compounds (Larralde et al. 1995). During the evolution of life on Earth, the carrier molecule of genetic information was changed at least once and RNA still plays a key role in modern biology. This indicates that such changes can occur without redesigning all of primitive proto-cell workings. It would thus seem that the way genetic information is stored is one flexible function in our hypothetic minimal life form, at least early in its formation.

6.1.4 Sources of energy

Today the most productive natural way of powering an organism is by either utilizing sunlight via photo-synthesis or by digesting something that does. While sunlight is the most productive energy source nowadays, there are a large number of alternatives. In fact it is very hard to find a place on Earth not inhabited by some sort of organism. Findings of very unusual organisms (Madigan and Marrs 1997) and references therein) in recent years seem to suggest that just about every chemical gradient imaginable can support some sort of extremophile life form.

On the other hand it is still a matter of scientific debate what energy source the first organisms on Earth made use of. A candidate for an early habitats are Black Smokers, liquid sulfur-rich hydrothermal vents on the sea floor that offer possibilities for energy production based on inorganic compounds (chemolithotrophic growth) (Miller and Bada 1988; Wächtershäuser 2000). These Black Smokers might not only have presented the energy source for early life, but might possibly also have provided catalytic functions assisting the polymerization of vital molecules (Wächtershäuser 1990).

6.2 How habitable are the considered habitats?

Another question which needs to be addressed in this context is the question of the time span necessary for the formation of a "living organism" from pre-biotic materials. This however is a question that can be answered only very vaguely and speculatively. The time necessary to produce a life from pre-biotic starting materials depends on a number of conditions, none of which can reliably be quantified at this time. In order to form a first proto-cell, a certain concentration of organic starting materials—the so called building blocks of life discussed in Sect. 2—need to be present. This critical concentration (which is not precisely known) determines the mass of organic material needed for a given volume of water. This volume could in principle range from a small shallow puddle to reactions occurring a vast planet-wide ocean (although the dilution effects of a large ocean make it less promising as a location for life's origins), giving very unspecific constraints on the concentration of organic chemicals needed.

On Earth the oldest reported microfossils seem to be from the Warrawoona group in Western Australia, which was formed about 3.5 billion years ago (Schopf and Packer 1987), although they are somewhat disputed (Brasier et al. 2005). Assuming that the

LHB phase, which has been proposed to have sterilized Earth, ended about 3.8 Gyr ago, this would mean that, on Earth, life started and evolved to sufficient complexity to leave behind a fossil record within the span of about 300 Myr. Of course that does not give us a date for the start of life on early Earth since the first organisms surely did not leave anything behind and it is very difficult to find rock formations on Earth that are that old, but at least it gives us some order of magnitude for the time span needed for life to emerge. It would thus seem reasonable to assume that several hundred million years are as good estimate to have originated life on Earth.

The four habitability classifications discussed in this review will now be discussed in terms of their potential to originate life and enable it to evolve.

6.2.1 Class I habitats: Earth-like planets

Habitable planets of Class I are a special case: we have only one example of such a planet, which is unfortunate, since the study of one planet alone yields results that are hard to generalize. This one case is, on the other hand, the best studied planet in existence, so there is a lot of data on the life it supports. Since this one planet is also the only one known to exhibit life, one cannot really say whether this is a statistical anomaly or a trait specific to all Class I habitable planets. All we can safely say is that such a planet is known to produce life and life could evolve from microorganisms to more complex life forms including humankind. How exactly life on Earth started and when is not entirely clear, but this is not because no-one can think of a reasonable scenario for the emergence of life on early Earth but rather due to the fact that there are so many different possible pathways to life that it is hard if not impossible to say which one is the one that life has actually taken (Nisbet and Sleep 2001). It would thus seem that on other planets exhibiting the same or similar conditions as early Earth the emergence of life should also be plausible.

6.2.2 Class II habitats: Mars/Venus-type planets

On Earth the emergence of life seems to have taken place almost immediately after the formation of a stable hydrosphere (Joyce 2002). If we assume that Class II planets do not yet differ from Earth that early in their evolution, it seems reasonable to assume that life could have started. Even if they do not have plate tectonics, early in their history there should have been enough active volcanoes (probably also under water) to provide local conditions comparable to those of the Black Smokers on Earth's sea floor. The question is, how long these conditions prevail until the planet evolves to be either too dry or too hot. If life has had enough time to evolve into sufficient complexity and to thrive in many different habitats, possibly utilizing a choice of different energy sources, it is very hard to eliminate it completely. This is shown by the fact that none of the mass-extinction events in Earth's history managed to extirpates life entirely (Raup and Sepkoski Jr. 1982). It seems possible that should life have started early in the history of Class II habitable planets and if favorable conditions prevailed long enough to allow for an evolution that it might persist even after the loss of (almost) all water. The production of complex and diverse ecosystems, however, depend on the carrying capacity of the planet and the circumstances how fast life may develop.

6.2.3 *Class III habitats: subsurface oceans with silicate contact on the sea-floor*

While in the cases of Class I and Class II habitable planets discussed above the main question is whether conditions favorable to life persist long enough to allow life to develop and evolve after it started, the question here is if life could start at all. This mainly has to do with the question of where the starting materials could come from. Assuming the organic material necessary to start life is supplied by impact of meteorites and comets and by precipitation of interplanetary dust, this material impacting on the surface has to find its way into the subsurface oceans. Also this material has to reach meaningful concentrations in some (small) compartment of the ocean, which is hard to imagine in a connected body of water as large as a planet-wide subsurface ocean. However, one should keep in mind that synthesis of organic material by either Fischer–Tropsch reactions or the kind of catalytic cycles described by Wächtershäuser (1990) are possible under the high pressure/high temperature conditions occurring at deep-sea vents. In such environments on Earth, reduced radicals such as H_2 are contained in the hot fluid and can provide energy for a variety of organisms. The source of energy necessary to power an organism could be another problem. Under such conditions it seems more difficult that life could start. Compared to class I and II habitats it seems more problematic for life originating in a subsurface ocean.

6.2.4 *Class IV habitats: subsurface oceans with no silicate-rich sea-floor and "Ocean planets"*

Class IV habitats and exoplanets where a water ocean is in contact with a thick ice layer ("Ocean planets"; Léger et al. 2004; Selsis et al. 2007) present a much better situation for the influx of organic material from outside the planet compared to bodies like the Jovian satellites Ganymede or Callisto. The main problem encountered in Class IV habitats is, however, much more severe: that of the concentration of the necessary ingredients for life. A planet whose surface is completely covered in water several kilometers deep with nothing to act as a concentrating "sponge" for organic chemistry is simply too vast for any two or more interesting molecules to meet. While a sea/ice system could in theory provide such a means of concentrating life's ingredients (Trinks et al. 2005), not even the starting concentrations needed for a system like that are likely to be reached in addition to the quite specific temperature conditions needed.

Considering all these four different classifications of habitable planetary bodies, it seems hard to imagine higher life forms as we know populating anything but a Class I habitable planet. If, however, life has had a sufficient head start and gained a foothold on a Class II habitable planet before it veers off its Class I course, it is imaginable that life can persist in some way on planets as cold and dry as Mars or as boiling as Venus, even though complex, or even multi-cellular organisms are most likely out of the question under such conditions.

7 Discussion

In the following pages we discuss how life or traces of life on these different habitats can be discovered and studied.

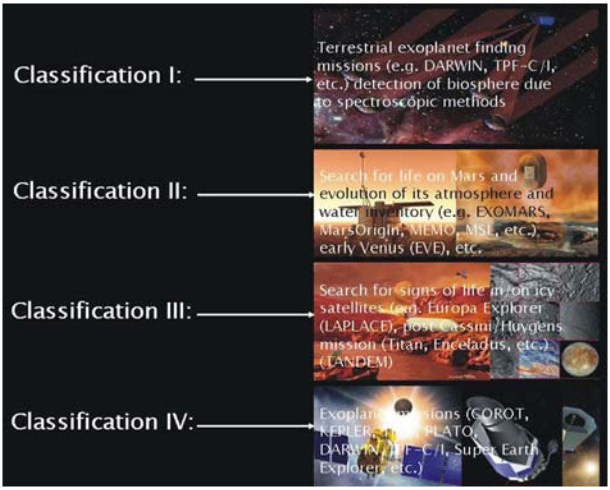


Fig. 21 Illustration of the habitats which we have discussed in this review article and possible space missions which will be designed for the detection of life or traces of life on such bodies

Figure 21 illustrates the various habitats and planned space missions which will be designed to detect life or traces of life in them. The discovery of life in different evolutionary stages on the four main habitat classes will also confirm or reject the habitability concept outlined and discussed in this review.

7.1 Detection and characterization of life on class I habitats

The detection of life on class I habitats on planets around other stars will be one of the greatest endeavors of mankind. Future missions like Darwin (ESA) (http://www.esa.int/esaSC/120382_index_0_m.html) and NASA's Terrestrial Planet Finder (http://planetquest.jpl.nasa.gov/TPF/tpf_index.cfm) Coronagraph/Interferometer are designed to detect biosignatures in exoplanetary atmospheres (e.g., Cockell et al. 2009). Biosignatures mean detectable atmospheric gas species, or a set of species, the presence of which at significant abundance strongly suggests a biological origin (DesMarais et al. 2002). This is the case for the couple $CH_4 + O_2$ or N_2O. The search for signs of life is based on the assumption that extraterrestrial life shares fundamental characteristics with life on Earth, in that it requires liquid water as a solvent and has a carbon-based chemistry. Life on the basis of a different chemistry is not considered here because the vast number of conceivable life-forms produces signatures in their atmosphere that are so far unknown. Therefore, we assume that extraterrestrial life on class I habitats may evolve more or less similar to life on Earth in its use of the same input and output gases, that it exists far from thermodynamic equilibrium, and that it has analogs to bacteria, plants, and animals on Earth. Their spectral signatures changed significantly on geological timescales on Earth (Kaltenegger and Selsis 2007).

The first step of exploration of terrestrial extrasolar planets will be a space mission that can detect and record-low resolution spectra of extrasolar planets. Detection of water vapor in a planetary atmosphere is the first clue indicating that a planet may be habitable. Water absorption can be seen in several bands in the visible and infrared making this compound easy to identify. On the visible-NIR (0.5–1 μm) the best regions to detect water absorption are 0.7, 0.8 and 0.9 μm. Other possibilities are the bands at 1.1, 1.7 and 1.9 μm. As for the mid-IR, water can be detected between 5 and 8 μm and from around 17 out to 50 μm. Methane (CH_4) and nitrous oxide (N_2O) absorb in the wavelength from 6 to 7 μm. With extremely high resolution, the absence of water can also be deduced from the presence of highly soluble compounds like SO_2 and H_2SO_4. Venus' spectrum, for example, has some weak absorption bands from water but its bulk atmospheric chemistry can only be explained if the planet is dry.

O_2, O_3, CH_4 are good biomarker candidates that can be detected by a low-resolution spectrograph (see, e.g., Kaltenegger and Selsis 2007; Segura and Kaltenegger 2008; DesMarais et al. 2002). These gases should be ubiquitous by-products of carbon-based biochemistry, even if the details of alien biochemistry are significantly different from the biochemistry on Earth.

Oxygen in high abundance is a promising bio-indicator, especially in combination with a reducing gas. A detectable concentration of O_2 and/or O_3 and of a reduced gas like CH_4 can be considered as a signature of biological activity. The spectrum of the Earth has exhibited a strong infrared signature of ozone at 9.6 μm for more than 2 billion years, and a strong visible signature of O_2 at 0.76 μm for a period of time between 2 and 0.8 billion years (depending on the required depth of the band for detection and also the evolution of the O_2 level) (Kaltenegger and Selsis 2007). The depth of the saturated O_3 band is determined by the temperature difference between the surface-clouds continuum and the ozone layer. CH_4 is not readily identified using low-resolution spectroscopy for present-day earth, but the methane feature at 7.66 μm in the IR is easily detectable at higher abundances, provided that the spectrum contains the whole band and a high enough signal to noise ratio. Depending on the degree of oxidation of a planet's crust and upper mantle nonbiological mechanisms can also produce large amounts of CH_4 under certain circumstances.

N_2O is another biomarker on Earth, detectable at 17.0 and 7.8 μm with high resolution. It would be hard to detect in Earth's atmosphere with low resolution, as its abundance is low at the surface (0.3 ppmv) and falls off rapidly in the stratosphere. N_2O is produced in abundance by life on Earth but only in trace amounts by natural processes. Nearly all of earth's N_2O is produced by the activities of anaerobic denitrifying bacteria. On a low-O_2 early Earth, its abundance would be even smaller because it photolyses rapidly in the near ultraviolet. There are other molecules that could, under some circumstances, act as excellent biomarkers, e.g., the manufactured chloro-fluorocarbons (CCl_2F_2 and CCl_3F) in our current atmosphere in the thermal infrared waveband, but their abundances are too low to be observed spectroscopically at low resolution (Segura et al. 2005).

As signs of life in themselves the greenhouse gases H_2O and CO_2 are secondary in importance because, although they are not indicators of its presence, they are raw materials for life and thus necessary for planetary habitability.

Planets that are more massive than Earth could potentially be habitable class I planets. Different groups have investigated the effect of different orbital distance, increased gravity and its changes of the temperature pressure profile, effects of cosmic rays and thus the chemical reaction rates for Earth-like super Earths (Grenfell et al. 2007, 2008; Kaltenegger et al. 2009). Biomarkers are detectable in the spectra of massive Earth analogs.

7.2 The search for life on class II habitats

Mars or Venus are interesting bodies to be studied for habitability in the Solar System. On both planets life may have once appeared and may have disappeared due to faster developing geological processes and changes. There is evidence that Mars once had a denser atmosphere, a warmer climate and an intrinsic magnetic field (e.g., Kulikov et al. 2007; Dehant et al. 2007; Lammer et al. 2008). On the other hand evidence exists that the Martian magnetic dynamo stopped before life may have developed and the atmosphere and the planet's water inventory were lost so that the planet went into a negative greenhouse cycle which resulted in the cold red planet we investigate today.

Mars is a small terrestrial planet which may be considered a kind of class II paleo-habitat and may help us in understanding how geological evolution and habitability are coupled. Furthermore, we may know if Mars ever hosted life and if so, how long and under which conditions Mars was able to sustain life (Langlais et al. 2009). As pointed out in a recent ESSC-EDF Position paper Science-Driven Scenario for Space Exploration (Worms et al. 2009), an answer to these questions is also crucial for understanding the habitability and limits for the evolution of life on exoplanets (http://www.esf.org). The search for extinct and extant life on Mars is ongoing through space missions like the Phoenix (NASA) lander, and future missions such as the NASA's Mars Science Laboratory (2011) and ESA's ExoMars mission, now scheduled for 2016. The long-term Mars exploration programs in the US and Europe will build on scientific discoveries from these missions which will hopefully lead to an international Mars sample return program.

The second potential class II paleo-habitat in the Solar System is Earth's inner neighbor Venus. Future Venus missions like the planned ESA/Russian European Venus Explorer (EVE) project (http://www.aero.jussieu.fr/EVE/) will have the capability for studying the complete record preserved in the elemental and isotopic composition especially that of noble gases in Venus' atmosphere, including quantifying the escape processes. Related to these tasks will be the study of the stability of the current Venus climate, by quantifying the exchange of atmospheric constituents with the surface and interior of the planet, and at the interface with space. By understanding these processes one can model the evolution of the climate of Venus, and the history of water and other volatiles. Other open questions are the cycling of water and sulphur compounds in the complex cloud environment. This will be addressed with simultaneous studies of the chemical, radiative, and dynamic processes from a Venus balloon. A detailed analysis of Venusian clouds will show if the clouds are maintained by a constant volcanic output of sulphur compounds into the atmosphere. Moreover, a study of the liquid (albeit highly acidic) water in the clouds may help us in understanding whether they can provide an environment where pre-biotic compounds or even microorganisms can survive.

The main task of future Venus missions, however, will be the re-construction of the geological history of the planet. This can be realized by mapping the structural elements of the shallow subsurface, so that volcanic episodes and formations can be better understood, and by exploring anomalies in the ionosphere that can be correlated to present subduction activities. All the tasks discussed before can only be carried out with the combination of orbiter, balloon and landers. As discussed in this review article, terrestrial planets within HZs of lower mass stars may be most likely evolving to Venus-type or Venus-Earth hybrid type planets instead of to Earth-analogs. The investigation of the evolution of Venus in comparison with the class I habitat Earth is therefore of fundamental importance for our understanding of the evolution of habitable planets in general.

7.3 Europa as a planetary object and potential class III habitat

Jovian's satellite Europa most likely represents the only class III habitat in the Solar System. This makes this body a special object. Due to the difficult direct access to the potential habitat, the in-situ exploration of subsurface oceans will be a long-term scientific endeavor. The planned Europa Jupiter System Mission (EJSM also named LAPLACE) a mission to Europa and the Jovian system will study, among many other science tasks the habitability of Europa in context of the Jupiter system and its links with the formation scenario and key coupling processes (http://jupiter-europa.cesr.fr/). The mission project has been submitted to ESA, but has merged with a NASA project. While an ESA spacecraft will mostly study Jupiter and Ganymede, a NASA orbiter will be devoted to Europa. For technical and budget reasons, it does not seem feasible to have on board the Europa orbiter a kind of surface element (e.g., penetrator) that could perform some in-situ characterization of the surface ice. However, there is a project for a true lander under study by the Russian Space Agency (Roskosmos). One of the science themes concerns the characterization of Europa as a planetary object and a potential habitat. The mission will probe the existence of Europa's subsurface ocean, establish its main characteristics and map the topography of its silicate sea-floor. It will study Europa's surface mineralogical and chemical composition, constrain its environmental parameters and look for possible bio-signatures. It will study the dynamics of the mantle, the ice crust and its coupling to the subsurface ocean. Finally, it will characterize Europa's exosphere and its interaction with Jupiter's magnetosphere and plasma environment.

7.4 Investigation of subsurface icy water worlds, Titan lakes and extra solar ocean planets

The Cassini-Huygens mission is a remarkable success, answering many outstanding questions about the Saturn system and Titan in particular. As for many successful missions, the key contributions of Cassini may be the questions raised rather than those answered. An important limitation of Cassini, with respect to Titan science, is the insufficient spatial coverage allowed by its orbit around Saturn. While measurements have highlighted the complexity of Titan's atmosphere and magnetic environment, the coverage has been insufficient to achieve a full understanding. The minimum possible flyby altitude of 950 km and the uneven latitudinal coverage have limited our ability to explore the full set of atmospheric chemical processes. Opportunities for occultations have been rare, thus gaps remain in the magnetospheric downstream region. The single vertical profile of the atmosphere taken by Huygens limits our understanding of horizontal transport and latitudinal variations.

The surface of Titan, as revealed by the Huygens probe and the Cassini orbiter, offers us an opportunity to stretch our current models in an effort to explain the presence of dunes, rivers, lakes, cryovolcanoes and mountains in a world where the rocks are composed of water ice rather than silicates and the liquid is methane or ethane rather than liquid water, but the limited high resolution spatial coverage (corresponding to 25–30% with radar and much less with VIMS) limits our view of the range of detailed geological processes ongoing on this body. The exciting results from the Huygens post-landing measurements are limited to a fixed site, short timescales, and do not allow for direct sub-surface access and sampling.

The two major themes in Titan exploration—the methane cycle as an analog to the terrestrial hydrological cycle and the complex chemical transformations of organic molecules in the atmosphere—make Titan a very high priority if we are to understand how volatile-rich worlds evolve and how organic chemistry and planetary evolution interact on large spatial and temporal scales. Both are of keen interest to planetology and astrobiology.

The intriguing discoveries of geological activity, excess warmth and outgassing on Enceladus (due perhaps to the ejection of water and organics from subsurface pockets bathed in heat, or by some other mechanism), mandate a follow-up investigation to that tiny Saturnian world that can only be achieved with high resolution remote observations, and detailed in situ investigations of the near-surface south polar environment.

The questions that will remain to be answered until a new mission to Titan and the Saturn system will be launched. The concept of such a future mission is currently being studied by ESA and NASA. The so called Titan Saturn System Mission (TSSM), is the merging of a Titan and Enceladus Mission (TandEM) proposal to ESA for the Cosmic Vision Plan (Coustenis et al. 2009) and the Titan Explorer 2007 Flagship study for NASA. The primary science goals of TSSM (http://www.lesia.obspm.fr/cosmicvision/tssm/tssm-public/) are to understand Titan's and Enceladus' atmospheres, surfaces and interiors, to determine the pre- and proto-biotic chemistry that may be occurring on both objects, and to derive constraints on the satellites' origin and evolution, both individually and in the context of the complex Saturnian system as a whole.

Many internal processes play crucial roles in the evolution of Titan and Enceladus. The formation and replenishing of Titan's atmosphere and the jet activity at Enceladus' South Pole are intimately linked to the satellite's interior structure and dynamics. Open issues are listed below:

- to determine their present-day structure and levels of activity;
- to determine whether the satellites underwent significant tidal deformation, and whether they possess intrinsic or induced magnetic fields and significant seismicity;

– to identify heat sources, internal reservoirs of volatiles (in particular methane and ammonia) and eruptive processes.

Besides the icy satellites of Solar System gas giants, class III/IV habitats may also be discovered by measuring the mean densities of discovered Super-Earth-type exoplanets. Selsis et al. (2007) studied this possibility for present day transit missions in space like CoRoT (CNES) and Kepler (NASA) in combination with ground-based Doppler velocimetry measurements HARPS (ESO) and possible future instruments. These authors studied the sources of uncertainty on the planetary density related to those on the mass determination by Radial Velocity measurements, the stellar radius determination, and the photometric measurement during the transits. They found that with the presently available instruments, the accuracy of Radial Velocity measurements is the main uncertainty and limiting factor for expected detections by CoRoT and Kepler.

As a consequence the determination of the nature of such planets seems only possible if they are in an adequate domain related to the host star magnitude (Selsis et al. 2007). Their study showed that if each star in the CoRoT field had a 6–10 Earth-mass planet at $\sim$0.10 AU, the number of detections with CoRoT would be several hundred. On the other hand the absence of the detection would indicate that such planets are not present at the level of $\sim$1%.

However, a definite answer to these open questions is related to the photometric precision of CoRoT and Kepler. New generations of Radial Velocity instruments are necessary, which can make accurate measurements on faint stars. In that case, the identification of Ocean Planets could be done on a significantly larger stellar sample.

8 Conclusions

The study of habitability indicates that plate tectonics, a global magnetic field, a hydrosphere and the distance of the HZ from its host star are crucial factors for planets to evolve to Earth-like, class I habitable planets. The stellar radiation and plasma environment to which slow rotating tidally locked terrestrial planets within the HZs of active low mass M and K stars are exposed strongly affects the evolution of their atmospheres and initial water inventories. Slow rotation which results in a weak magnetic field coupled with the proximity to their host star could result in the loss of most of their initial atmosphere. The planetary atmosphere would have to be replenished until the stellar activity is low enough to maintain the planetary atmosphere and water inventory.

For Class II habitats the crucial point is related to the timescale of favorable Class I-type environmental conditions which allow for the origin and evolution of life.

Life on habitable planets of class III and IV may only be detected through in situ measurements. Life on exoplanets can only be detected remotely if it influences its atmosphere in detectable ways; therefore, we expect that class II–IV habitats can most likely be explored in our own Solar System.

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