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Water, Life, and Planetary Geodynamical Evolution

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Abstract In our search for life on other planets over the past decades, we have come to understand that the solid terrestrial planets provide much more than merely a substrate on which life may develop. Large-scale exchange of heat and volatile species between planetary interiors and hydrospheres/atmospheres, as well as the presence of a magnetic field, are

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important factors contributing to the habitability of a planet. This chapter reviews these processes, their mutual interactions, and the role life plays in regulating or modulating them.

Keywords Mantle dynamics · Habitability · Magnetic field · Volatiles · Thermal evolution

1 Introduction

Although in our current knowledge life appears to be limited to a relatively narrow and superficial zone extending no more than a few kilometers from Earth's surface in either direction, the vast volume of mostly solid rock and solid or liquid iron alloy which is inside this envelope provides several functions that are important if not crucial for planetary habitability. It appears logical that these functions may also contribute to the habitability of other planets (or their absence to hostility towards life). Life has not been detected on any other astronomical body, but there have been inferences and speculations for Mars (e.g. Formisano et al. 2004) and Europa (see Chyba and Phillips 2001). Titan has also attracted the attention of exobiologists (Raulin and Owen 2002).

This chapter aims to provide an overview of the many processes and interactions of the interior of terrestrial planets which may help sustain life. Obviously this is a field of study which is only starting to be uncovered, and on many subjects our knowledge and understanding is far from complete. Therefore we also try to discuss some important open questions which need to be answered.

Figure 1 gives a general overview of the direct and indirect interactions of planetary core, mantle, crust, hydrosphere/cryosphere and atmosphere processes with the biosphere for planets with plate tectonics (Earth, and possibly early Mars and early Venus) and for one-plate planets (Mars, Venus). The transfer of heat from the core through the mantle and out into the atmosphere and space is shared by both types (though in a one-plate planet, the core and the mantle are not necessarily cooling down), as are degassing and heat transfer by volcanism. Both show erosion of the atmosphere by the solar wind, but the presence of a magnetic field may mitigate this in the case of plate tectonics. However, only the plate tectonics scenario includes a means of reintroducing material back into the planet's interior through subduction, generating a cycle between interior and exterior reservoirs, though a mechanism has been proposed for one-plate planets as well (Elkins-Tanton 2007a).

It is immediately clear that several interactions consist of the transfer of volatile species like water and CO_2 between different reservoirs. This is very important since both species are the essential materials of life (nitrogen, phosphorus and sulfur are other important basic raw materials), and also main constituents of the hydrosphere/cryosphere and atmosphere.

Less obvious, but not less important, is the effect that volatiles, and specifically water, may have on the dynamics of the planetary interior. The presence of water is thought to play a crucial role in the generation of plate tectonics through weakening the lithosphere (Regenauer-Lieb et al. 2001). The presence or absence of plate tectonics significantly affects the thermal evolution of a planet, which in turn determines whether a magnetic field can be generated to protect the atmosphere against the solar wind (see Chap. 6). In addition to this, volatile species play a significant role in determining the surface temperature of a planet, thus affecting the temperature difference between interior and exterior which drives the dynamics of the mantle.

In this chapter, these interactions are reviewed in more detail, combining two approaches to habitability. On the one hand, we stay on firm ground by looking at Earth and what makes it habitable. On the other hand, we look at other planets, specifically Mars and Venus, and

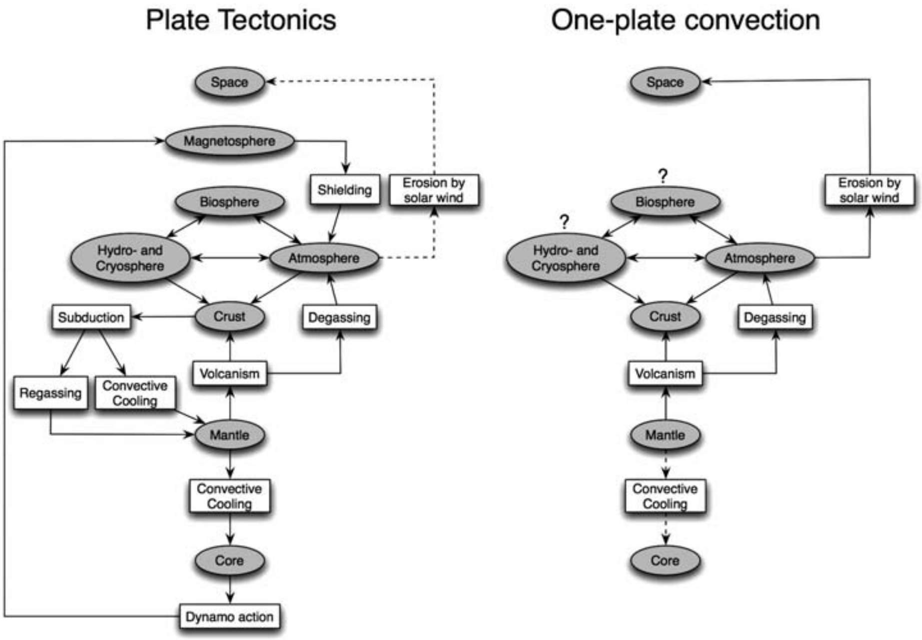

Fig. 1 Sketch comparing the possible relationships between interior, surface and atmosphere of a planet with plate tectonics on the one hand, and a planet with convection underneath a single plate on the other hand. *Dashed arrows* denote low efficiency mechanisms. By promoting very complex and global geological cycles, plate tectonics tends to generate conditions favorable for life. In the case of a one-plate planet (no subduction) the geologic interactions between the different subsets are reduced and more localized. Such a planet may not be able to sustain its internal magnetic field and recycle volatiles (extracted from the interior by magmatism) into the mantle; finally it evolves toward a planet with either an atmosphere that is too thin (Mars) or too thick (Venus), being quite probably uninhabitable at the present time

consider what makes these planets less suitable for life. We start by discussing the inventories and fluxes of volatiles in terrestrial planets (focusing on Earth, which is the only planet for which we have sufficient information) in Sect. 2. In Sect. 3, we discuss several processes involving these volatile species which are important for maintaining planetary habitability. The dynamical evolution of Earth, Mars and Venus in the context of these processes is presented in Sect. 4, and finally a synthesis discussing the implications for planetary habitability and open questions is presented in Sect. 5.

2 Origins, Inventories and Cycles

In the light of the central position volatiles appear to play in the direct and indirect interaction between planetary interiors and biospheres, the following sections discuss the present-day inventories and fluxes of volatiles in Earth's interior and exterior and the Martian and Venusian atmospheres, and the origin and survival of these inventories. The focus will be on H_2O and CO_2, as these appear to be the most important and best studied. As N and S are crucial components of amino-acids, these elements will be treated as well. We finish by discussing the oxidation state of the mantle and its implications for habitability.

2.1 Earth's Volatile Inventories

Even for planet Earth, which is the best studied of the terrestrial planets, there remains considerable uncertainty about distribution of inventories between different reservoirs, the core remaining the most elusive. Table 1 shows a compilation of recent reservoir size estimates for Earth's core, mantle, and hydrosphere/atmosphere.

Most mantle minerals, including nominally anhydrous minerals, may carry several hundred ppm's of water. Xenolith samples have shown the following ranges of water contents for upper mantle minerals (Withers et al. 1998; Ingrin and Skogby 2000; Koga et al. 2003; Mosenfelder et al. 2006): 100–1300 ppm in clinopyroxene, 60–650 ppm in orthopyroxene, up to several thousand ppm in olivine, and up to 1000 ppm in garnet. In the transition zone, $(Mg,Fe)_2SiO_4$—γ-spinel can contain up to 2.7 wt.% of H_2O (Inoue et al. 1998; Williams and Hemley 2001). This may make the transition zone an important reservoir for the Earth's interior water (Bercovici and Karato 2003). Mg-perovskite may contain up to 700 atoms of H per 10^6 atoms of Si at 27 GPa and 1830°C, but increasing the pressure and adding Al may increase this solubility (Williams and Hemley 2001). Murakami et al. (2002) report that magnesiowüstite and $MgSiO_4$-rich perovskite may contain 0.2 wt.% water, and $CaSiO_4$-rich perovskite may contain 0.4 wt.%. These numbers suggest that the lower mantle may contain up to 5 ocean masses. The Earth's core may be the most important of the hydrogen reservoirs (Williams and Hemley 2001; Hirao et al. 2004), as shown in Table 1, storing hydrogen in the form of iron hydrides. High pressure experiments of Saxena et al. (2004) confirm the stability of iron hydride at core pressures.

Most of the Earth's carbon is in its interior. However, the solubility of CO_2 in olivine, pyroxenes, garnet and spinel is very low, about 0.1–1 ppm by weight. Keppler et al. (2003) conclude from this that there must be a separate carbon-bearing phase in the deep upper mantle, possibly carbonates. The carbon content of the atmosphere and oceans is orders of magnitude smaller. The Earth's core may contain 2–4 wt.% C (Wood 1993). Core compo-

Table 1 Inventories of important volatile species in the Earth's interior and exterior, and the martian and venusian atmospheres

Element/compound	Reservoir size (g)				
	Earth			Mars atmosphere[o]	Venus atmosphere[p]
	Core	Mantle/crust	Hydrosphere/atmosphere		
H_2O	$1 \cdot 10^{25}$ – $1.4 \cdot 10^{26}$ [a,b,n]	$3.6 \cdot 10^{24}$ – $8.3 \cdot 10^{24}$ [a,n]	$1.4 \cdot 10^{24}$ [m]	$2.3 \cdot 10^{15}$	$4.8 \cdot 10^{18}$
C	$4 \cdot 10^{24}$ – $8 \cdot 10^{25}$ [e,n]	$3 \cdot 10^{22}$ – $4.9 \cdot 10^{23}$ [c,d,n]	$3.6 \cdot 10^{19}$ [l]	$6.5 \cdot 10^{18}$	$1.3 \cdot 10^{23}$
N	$1.4 \cdot 10^{23}$ [n]	$3.8 \cdot 10^{20}$ – $8.1 \cdot 10^{21}$ [d,h,n]	$4 \cdot 10^{21}$ [f]	$4.4 \cdot 10^{17}$	$1.1 \cdot 10^{22}$
S	$1.9 \cdot 10^{25}$ – $3.3 \cdot 10^{25}$ [g,i,n]	$1.0 \cdot 10^{24}$ [h,n]	$1.3 \cdot 10^{21}$ [k]		$5.3 \cdot 10^{19}$

References: a) Williams and Hemley (2001), b) Hirao et al. (2004), c) Coltice et al. (2004), d) Zhang and Zindler (1993), e) Wood (1993), f) Galloway (2001), g) Allègre et al. (1995), h) McDonough and Sun (1995), i) Dreibus and Palme (1996), j) Jahnke (1992), k) Charlson et al. (1992), l) Holmen (1992), m) Murray (1992), n) McDonough (2003), o) Williams (2004), p) Williams (2005) sitions of 0.3 wt.% carbon or more, in the presence of another light element like S, results in the formation of Fe_3C as the first crystallizing phase, which is more consistent with the inner core density than a pure Fe or FeNi composition.

It is unclear whether the core reservoir has significant interaction with the mantle, and thus ultimately with the biosphere. Tungsten isotope data have been applied to infer the presence or absence of core material in the source material of oceanic island basalts, but the results are equivocal and remain controversial (Scherstén et al. 2004; Jellinek and Manga 2004; Humayun et al. 2004), with both geochemical and geophysical evidence arguing against it (Lassiter 2006).

2.2 Volatile Inventories of Mars and Venus

Obviously, the volatile inventories of Mars and Venus are much more difficult to estimate than those for Earth. Most information is available on the atmospheres, which is listed in Table 1. Note that on Mars, significant exchange (20% of atmosphere mass) of volatiles between the atmosphere and the surface, the seasonal ice caps, and the polar layered deposits takes place on a seasonal basis (Mischna et al. 2003; Aharonson et al. 2004). The north polar cap primarily consists of water, and has an estimated permanent water inventory of $1.4 \cdot 10^{21}$ g (Zuber et al. 1998). The south polar cap may contain, when assuming it to consist of water only, about $2.5 \cdot 10^{21}$ g (Smith et al. 1999). In addition to this, there appear to be significant volumes of ice in the Martian subsurface (Mitrofanov et al. 2002; Feldman et al. 2002; Boynton et al. 2002).

Dreibus and Wänke (1987) estimated the water content of the Martian mantle to be 36 ppm, i.e. a relatively dry mantle. This corresponds to a mantle inventory on the order of 10^{22} g, or two orders of magnitude smaller than in Earth's mantle (see Table 1). Note however that the uncertainty in this number is significant (see Lodders and Fegley 1997), and a homogeneous distribution in the Martian mantle is not to be expected. To illustrate this point, we mention that Dann et al. (2001) and McSween et al. (2001) infer pre-eruptive water contents of Shergottite magmas of 1.8%, though this is disputed by Jones (2004). The Venusian mantle is generally assumed to be dry (see Nimmo and McKenzie 1998) like its atmosphere (Grinspoon 1993).

2.3 Origin and Survival of the Volatile Inventories

Oceanic island basalts samples around the globe show a large range of $^3He/^4He$ ratios. Whereas the latter of the two helium isotopes is the product of α-decay of several radioactive elements, the former is not. Since noble gases do not easily bind to anything and, as a consequence, there is nearly no subduction of noble gases (Staudacher and Allègre 1988), the presence of ^3He in the mantle suggests that ^3He was added to our growing planet during accretion, and that some of this primordial reservoir is still present. Similarly, the uniformity of the isotope composition of C in different types of basalts but also diamonds and carbonitites has been interpreted as evidence for a primitive C source in the mantle (Tingle 1998).

A relatively small amount of water in the original building blocks of the terrestrial planets, which could later have progressively degassed, would suffice to explain the present-day water budget of planet Earth. But contrary to carbon, there has been significant escape of hydrogen to space (see below). This may demand additional post-accretional sources. Possible sources are (see Gaidos et al. 2005 and references therein): (1) material from e.g. asteroids;

(2) comets; (3) ice particles from the outer region of the primordial solar system; (4) 'wet' planetary embryos.

During core segregation, significant amounts of hydrogen may have been transported into the core resulting from reactions between liquid metallic iron and water in the mantle (Ohtani et al. 2005):

$$3Fe + H_2O \rightarrow 2FeH_x + FeO. \tag{1}$$

Ohtani et al. observed that the reaction can take place between 6 and 84 GPa, thus being able to scavenge water from a large part of the primordial mantle. Okuchi (1997) suggests that more than 95 mole% of the water present in a primordial terrestrial magma ocean may have reacted for form FeH_x and moved into the core.

SNC meteorite studies by Dreibus and Wänke (1987) suggest that the Martian mantle is very poor in water, containing about 36 ppm. This appears to be in conflict with the general trend of increasing water availability which is observed when moving away from the sun in the solar system. However, other volatile elements like Na and K are more abundant in Mars than in Earth (references in Kuramoto 1997). It is likely that a lot of water was lost from Mars by the oxidation of metallic iron in the Martian mantle:

$$H_2O + Fe \rightarrow FeO + H_2. \tag{2}$$

The hydrogen which is released in this way can either dissolve into the core (Zharkov and Gudkova 2000), or be segregated into the atmosphere, from where it escapes into space (Hunten et al. 1987). However, an equilibrium may already have been reached during the accretion process in the planetesimals, possibly eliminating further loss after accretion.

Kuramoto (1997) proposed an alternative reducing agent: reduced C species. He argues that, whereas the Earth's core may contain significant amounts of carbon, the high sulfur content of the Martian core may have prevented the uptake of large amounts of C into the core, leaving it in the mantle. The idea of the core as a C reservoir is supported by the fact that graphite and carbites are common in iron meteorites, the geochemistry of these minerals suggests a planetesimal core origin of these meteorites, and the carbon isotope composition is similar to that of Earth's mantle (Tingle 1998, and references therein). Furthermore, Wood (1993) has shown that C is quite soluble in iron liquids at core pressure, and that dissolution of C in iron liquids during planetesimal differentiation may be expected.

2.4 Earth Volatile Cycles

Table 2 summarizes the fluxes of the species listed in Table 1 for Earth.

2.4.1 Volatile Fluxes

The subduction system as a whole is an important agent in the recycling of volatile species into the mantle. However, significant fractions of these species may not enter the deep mantle but may be reinjected into the atmosphere by arc volcanism.

Water is carried into the mantle in the form of sediment and igneous rock pore and structural water (Jarrard 2002). The most important water carrying minerals in subducting crust are probably lawsonite ($CaAl_2Si_2O_7(OH)_2 \cdot H_2O$, 11 wt.% water) and serpentine ($Mg_3Si_2O_5(OH)_4$, up to 13 wt.% water) (Williams and Hemley 2001). Serpentine is stable up to halfway down the upper mantle, and lawsonite down to the deep upper mantle. Water-carrying phases of lesser or unknown importance are zoisite ($CaAl_3Si_3O_{12}(OH)$, significantly lower water content), magnesian pumpellyite ($Mg_5Al_5Si_6O_{21}(OH)_7$), K-amphibole, alkali-rich hydrous phases, humite ($nMg_2SiO_4 \cdot Mg(OH,F)_2$), and Dense Hydrous Magnesium Silicate, or the so-called alphabet phases (Williams and Hemley 2001). Some of these minerals are stable down to great pressures (e.g. DHMS phase B 12–24 GPa, DHMS phase D down to 50 GPa) and may play a role in transporting subducted hydrogen down into the deep mantle. Apart from these mafic minerals, sediments also appear to play a role in transporting volatiles into the mantle. Plank and Langmuir (1998) calculated the composition of the average subducted sediment, and from this estimated global subduction fluxes of sediments and their components. Their estimate for the flux of H_2O carried by sediments during subduction is $9.49 \cdot 10^{13}$ g/yr, which is about 1/9 of the total H_2O subduction flux.

The fraction of the amount of subducted water which is released through arc volcanism is sometimes called the recycling ratio R_{rec} (Franck and Bounama 2001). Its complementary value, i.e. the fraction of water which is actually subducted into the deep mantle is usually expressed as the regassing ration R_{H_2O}. The geophysical modeling of Franck and Bounama (2001) gives a preferred value for R_{H_2O} of 0.3 to 0.4. Numerical mantle convection models confirm that parts of the subducting slab remain sufficiently cool to prevent complete dehydration (Van Keken et al. 2002). Rüpke et al. (2004) found that specifically serpentinized lithospheric mantle may be an important agent in transporting up to 40% of it's initial water content into the mantle, corresponding to $R = 0.6$. In contrast, geochemical based estimates for R_{H_2O} are much higher, around 0.9 (e.g. Ito et al. 1983), though the value of Peacock (1990) of about 0.16 is at the other end of the spectrum.

A common way to estimate the flux of CO_2 at mid-ocean ridges is by determining volatile contents of glasses sampled at mid-ocean ridges and calibrating C to ^{3}He (Marty and Tolstikhin 1998). Alternatively, pre-eruptive volatile contents of normal mid-ocean ridge (MOR) basalts can be determined (Saal et al. 2002), which results in similar flux estimates. Aubaud et al. (2004) looked at kinetic desequilibrium during degassing of MOR magmas, and reconstructed the original carbon content of the magma. They extrapolated this value to the global ridge system and estimated a carbon flux of $1.6^{+0.6}_{-0.3} \cdot 10^{14}$ g/yr of carbon, which is equivalent to $5.9^{+2}_{-0.7} \cdot 10^{14}$ g of CO_2 per year, i.e. much higher than previous estimates.

Carbon is subducted mostly in the form of carbonates added within the basaltic crust through hydrothermal circulation (Jarrard 2002) and carbonate sediments deposited at the top of the oceanic crustal column (Plank and Langmuir 1998; Kerrick and Connolly 2001). Plank and Langmuir (1998) report carbonate sequence thicknesses of up to several hundred meters at subduction zones in the subducting crust. Because of a limited number of direct measurements of arc volcanic CO_2 fluxes and uncertainty about representivity of present-day fluxes over geological history, it is better to estimate carbon fluxes based on magma production rates (Marty and Tolstikhin 1998). This requires knowledge of relative amounts of slab-derived and mantle-derived carbon. Based on $CO_2/^3He$ ratios and carbon isotopes in arc gases, Varekamp et al. (1992) concluded that more than 80% of arc volcanic CO_2 originates from subducted carbon species. Kerrick and Connolly (2001) studied metamorphic devolatilization of sediments and found that release of CO_2 requires specific conditions of pressure and temperature and also the presence of H_2O rich fluids, which explains the large amount of carbon that is not released.

Nitrogen is subducted in the form of ammonium and nitrates in the sediment cover of the oceanic crust. Ammonium is produced by life, and although nitrate can be produced inorganically by lightning, this requires the presence of oxygen, which in turn is produced by photosynthetic life (Boyd 2001). Much of this sedimentary nitrogen is returned into the atmosphere through arc volcanism. Thus, about 85% of arc volcanic nitrogen is of sedimentary origin (Sano et al. 2001a). The total net (i.e. atmospheric nitrogen subtracted) nitrogen subduction zone flux is $6.0 \cdot 10^8$ mol/yr, including both arc volcanism and backarc spreading centers (Sano et al. 2001b).

Sulphur is fixated in the oceanic crust and its sediment cover in the form of metal sulfides like pyrite. Thus, these are subducted into the mantle with the downgoing slab. Additionally, sulfates in pore water in the subducting slab and occasional evaporitic sulfate deposits may be subducted (Canfield 2004).

2.4.2 Non-steady State Contributions

Although Table 2 shows the present-day fluxes which can be considered to represent most of the past 180 My, during which the plate creation and subduction rates were probably relatively constant (Rowley 2002), there are punctuated events which form a large additional source of volatiles: the eruption of flood basalts. For example, the eruption of the Deccan flood basalts, around 65 Myr ago, resulted in the eruption of about $6 \cdot 10^{18}$ g carbon into the atmosphere over a period of 1.36 Myr (Wignall 2001), corresponding to an additional flux of $4 \cdot 10^{12}$ g/yr, which is similar to the lower end of the estimates for the present-day mid-ocean ridge flux. In addition to this, and more importantly, about $6 \cdot 10^{18}$ g of S was also injected into the atmosphere, also corresponding to an annual flux of $4 \cdot 10^{12}$ g/yr (Wignall 2001), which is significantly more than the present-day flux of arcs, ridges and hotspots combined. The magmas associated with a flood basalt event are not only a direct source but also an indirect source of carbon, since interaction of voluminous melts with carbon bearing sediments in the crust may cause additional release of carbon into the atmosphere (e.g. Svensen et al. 2004). On the early Earth, flood basalt eruptions were both larger and more frequent (Abbott and Isley 2002). Therefore, they were probably a more important factor in the atmosphere/hydrosphere volatile balance.

2.4.3 Flux Balances

Rüpke et al. (2004) investigated the evolution of subduction zone water transport in the course of Earth's history. Assuming a high spreading rate and young average age of the oceanic crust for the early Earth, and the inverse for the present-day Earth, they use a simple model to predict a high rate of degassing and low rate of reinjection of water in the early Earth. A flipping of the balance between degassing and regassing would take place somewhere between 2.5 Ga and 900 Ma, depending on assumptions of the amount of serpentinization of oceanic lithospheric mantle. Since then, the total amount of hydrosphere and atmosphere water has been slowly declining at the expense of mantle water. The models of Rüpke et al. (2004) predict a sea level drop of several hundred meters over the last 600 Myr.

Edmond and Huh (2003) argue that at the present day there is an imbalance between deposition of CO_2 in the form of $CaCO_3$ at ocean bottoms and subduction of this material, as the Atlantic and Indian oceans form significant regions of $CaCO_3$ deposition but have no significant subduction of this material. This illustrates the possibility of changes in the carbon cycle on time scales of the Wilson cycle (periodic opening and closing of ocean basins), and a weak direct coupling between rates of CO_2 consumption by weathering and recycling of fixated carbon through subduction zone degassing (Edmond and Huh 2003).

For carbon, nitrogen, and sulfur, Table 2 shows a net flux out of the mantle. The sulfur flux has probably been positive for the past 700 Myr, due to the oxic nature of the oceans. A significantly larger subduction flux of sulfur has been inferred for the Archean and Proterozoic, when the ocean water was anoxic, allowing massive deposition and subduction of pyrite (Canfield 2004).

Table 2 Fluxes of important volatile species between the Earth's interior and exterior

Element/compound	Fluxes (g/yr)				Balance
	Subduction ↓	Arc and backarc ↑	MOR ↑	Hotspots ↑	
H_2O	$8.7 \cdot 10^{14}$– $1.83 \cdot 10^{15}$ [a,n]	$1.4 \cdot 10^{14}$ [a]	o	o	↓ [j,k]
C	$1.1 \cdot 10^{13}$– $4.2 \cdot 10^{13}$ [b,d,e,n]	$3.0 \cdot 10^{13}$ [c]	$0.6 \cdot 10^{13}$– $2.2 \cdot 10^{14}$ [c,g,h,i]	$3.5 \cdot 10^{11}$– $4 \cdot 10^{13}$ [c,h]	↑
N	$9.0 \cdot 10^{9}$ [f]	$1.8 \cdot 10^{10}$ [f]	$1.1 \cdot 10^{10}$ [f]	$6.2 \cdot 10^{7}$ [f]	↑
S	$1.2 \cdot 10^{13}$– $2.4 \cdot 10^{13}$ [m]	$2.8 \cdot 10^{13}$ [l]			↑

References: a) Peacock (1990), b) Varekamp et al. (1992), c) Marty and Tolstikhin (1998), d) Plank and Langmuir (1998), e) Kerrick and Connolly (2001), f) Sano et al. (2001b), g) Resing et al. (2004), h) Sano and Williams (1996), i) Aubaud et al. (2004), j) Rüpke et al. (2004), k) Wallmann (2001), l) Charlson et al. (1992), m) Canfield (2004), n) Jarrard (2002), o) no estimate found in the literature

2.5 Volatiles and the Oxidation State of Earth's Mantle and Surface

As all life is driven by redox gradients, and significant exchange of chemical species between planetary interiors and atmospheres may take place, it is important to consider the mantle oxidation state when talking about habitability.

The initial oxidation state of a planetary mantle is probably already determined during accretion and early magma ocean stages. One possibly important process is the oxidation of ferrous iron by hydrogen with subsequent escape of hydrogen into space or sequestration in the core (Hunten et al. 1987; Ohtani et al. 2005), though this is difficult to quantify, and fails to explain the relatively reduced state of Martian basalts where this planet should be more volatile-rich than Earth (Wade and Wood 2005). Wade and Wood (2005) proposed that a layer of magnesium silicate perovskite may act as an 'oxygen pump' during accretion. In the perovskite, $Mg^{2+}Si^{4+}$ may be substituted by $Fe^{3+}Al^{3+}$ during fractional crystallization at the bottom of a deep magma ocean. This substitution is independent of the redox state and a disproportionation may take place of the form $3Fe^{2+} = 2Fe^{3+} + Fe^0$ (Frost et al. 2004). As the metallic iron which is formed in this way may be transported into the core by sinking iron diapirs, and exchange of silicate material between the perovskite and the magma ocean (and later the overlying peridotite) may take place, this results in the effective oxidation of the mantle. As Mars is probably not big enough to have significant amounts of perovskite in its mantle (the pressure in the lowermost mantle barely reaching that of the phase transition), this may also explain the apparently relatively reduced state of its mantle (Wade and Wood 2005).

The geochemistry of old and recent rocks may provide evidence for the evolution of the redox state of the Earth's mantle. A suitable set of tracers is V and Sc, since these elements behave in a similar fashion during partial melting of the mantle, and only V is sensitive to redox conditions during a differentiation event (Li and Lee 2004). Recently, iron isotopes ($^{57}Fe/^{54}Fe$) were proposed as a proxy for the mantle oxidation state, since ^{57}Fe preferentially fractionates into bonds with Fe^{3+}, whereas ^{54}Fe prefers to associate itself with Fe^{2+} (Williams et al. 2004). The heterogeneity of this isotope ratio in the mantle is larger than can be explained by metasomatism and partial melting. Therefore, this suggests that part of the variation is caused by a secular variation in the mantle oxidation state (Williams et al. 2005). However, other workers, using combinations of V, Cr and/or Sc as a proxy for redox state, find that the mantle oxygen fugacity has been relatively constant since the Archean (Canil 1997; Li and Lee 2004), possibly since ∼4 Ga (Delano 2001). The upper mantle may have undergone a progressive oxidation due to an imbalance between the effects of crustal recycling and introduction of carbon from the deep mantle (Kadik 1997), but so far this appears to be ill constrained by the available data.

The advent of an oxygen-rich atmosphere, also called the Great Oxidation Event (GOE, around 2.3 Ga, Holland 2002), has obviously had an enormous influence on the evolution of life on Earth. The relative constancy of the mantle oxidation state before and after this event suggests that the great oxidation event was not related to changes in the redox state of the mantle (Li and Lee 2004). However, Holland (2002) states that the GOE may have been related to "a gradual increase in the oxidation state of volcanic gases and hydrothermal inputs". A coupled change in the the oxidation states of the mantle and atmosphere was originally proposed by Kasting et al. (1993). Holland (2002) determined that the required change in the oxidation state is sufficiently small to be indiscernible in the present record of Cr and V data. Kump et al. (2001) proposed the following scenario: During the pre-GOE era, reaction of water with ferric iron in basalts produces hydrogen, which may escape into space, and ferrous iron. The operation of this process and subsequent subduction of oxidized crust into the deep mantle over a long period of time results in the formation of an oxidized reservoir in the deep mantle. During two periods of intense volcanism in the late Archean and early Proterozoic, the oxidized reservoir provides source material for the volcanism, and produces more oxidized associated volcanic gasses. This change in the oxidation state of (reduced) volcanic gasses, which form an atmospheric oxygen sink, could have changed the balance with photosynthetic oxygen production (cyanobacteria existed long before the GOE, see e.g. Brocks et al. 1999), resulting in an oxidizing instead of a reducing atmosphere (Kump et al. 2001).

Although there are several uncertain or even speculative aspects to this scenario, it does illustrate the possible significance of endogenic processes for the oxidation state at a planet's surface and therefore the conditions under which life may form and evolve.

3 Mechanisms and Feedbacks

3.1 Geodynamical Regimes

As will be discussed below, several processes characteristic of one or more geodyamical regimes play significant roles in creating and maintaining the habitability of a planet. Before discussing these, we first present an overview of geodynamical regimes.

3.1.1 Overview

Four distinct geodynamical regimes can be distinguished, three of which are depicted explicitly in Fig. 2, and the fourth, an episodic switching between two of the former regimes, implicitly. The most obvious is plate tectonics, which is active on present-day Earth. From an instantaneous point of view, characteristics of plate tectonics include spreading ridges where oceanic crust is created, and generally subparallel to these ridges, subduction zones where oceanic lithosphere is returned into the mantle. On longer time scales, the opening, growth, shrinking and closure of oceanic basins (Wilson cycles) have been inferred (Wilson 1966;

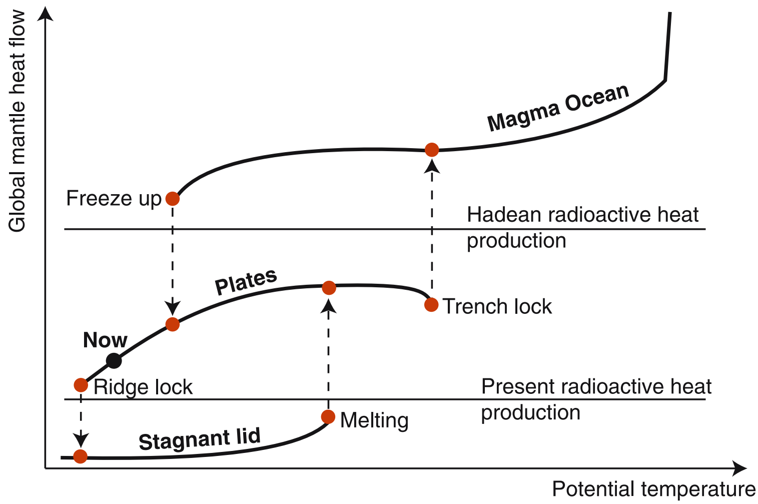

Fig. 2 Different branches of the relation between heat flow and temperature corresponding to different regimes in which the Earth and any other terrestrial planet could have operated. Adapted from Sleep (2000)

Turcotte and Schubert 2002). The main driving force of plate tectonics is *slab pull*, which is exerted by the subducting slab. In general, this pull is transmitted through the slab in its shallower parts, whereas deeper parts tend to exert their contribution through viscous drag of the mantle (Conrad and Lithgow-Bertelloni 2002). Mid-ocean ridge topography exerts a *ridge push*, which also contributes to the driving force (Turcotte and Schubert 2002). Additionally, *trench suction* associated with arc rollback (e.g. Wilson 1993) may play a significant role. The main opposing forces are lithosphere buoyancy (Oxburgh and Parmentier 1977; Sleep and Windley 1982; Vlaar and Van den Berg 1991), depending on the age of the slab and mantle temperature, and bending forces at the subduction zone (Conrad and Hager 2001).

The stagnant lid regime is well known from numerical and laboratory studies of fluids with a strongly temperature dependent viscosity (e.g. Davaille and Jaupart 1993, 1994; Reese et al. 1998), which results in a thick *lid* being formed on top of a convecting fluid layer. The efficiency of heat transport of this mechanism is significantly smaller than that of plate tectonics. Mars is thought to be in the stagnant lid regime (Spohn et al. 2001), showing a very old surface (Zuber 2001) with no evidence of subduction (though the hypothesis of an early phase of plate tectonics has been proposed, Sleep 1994; Connerney et al. 1999, 2005).

An episodic regime, with alternating phases of lid stagnation and lid mobilization, has been observed in numerical convection models (e.g. Stein et al. 2004; Van Thienen et al. 2004b). It has been suggested that Venus may presently be in this mode of convection (Parmentier and Hess 1992; Turcotte 1995), since crater counting studies indicate a relatively uniform surface age between about 300 Ma and 1 Ga (Schaber et al. 1992; McKinnon et al. 1997), which suggests a period of large scale resurfacing followed by a period of quiescence.

During the earliest phases of planetary evolution, sufficient energy was probably available to melt entire sections of planetary mantles, resulting in the magma ocean regime. The main sources of energy were the release of gravitational energy during accretion and core segregation (Horedt 1980), and the decay of short-lived and longer-lived radioactive isotopes. Some large impacts during the final stages of accretion were also sufficiently energetic to produce a magma ocean. Abe (1997) computed that the survival time of magma oceans generated in this way are on the order of 100–200 Myr for Earth.

Systematic studies of the conditions under which stagnant lid convection, active lid convection, and the intermediate episodic regime take place have been done by several workers (Solomatov 1995; Moresi 1998; Kameyama and Ogawa 2000; Stein et al. 2004). In general, the balance between the viscous and/or brittle strength of the lithosphere and convective stresses in the mantle determines the style of convection. In these numerical studies, a stagnant lid regime is found for situations with a high lithospheric yield stress and relatively low vigor of convection (Rayleigh number), and for high viscosity contrasts between the lid and the interior. A mobile lid regime is observed for a relatively low viscosity contrast, high Rayleigh number, and relatively small yield stress. Parameters for the transitional episodic regime are between these extremes.

3.1.2 Role of Volatiles in the Operation of Geodynamic Regimes

Volatiles, principally water, affect the operation of any regime primarily due to their weakening effect on the rheology of mantle rocks (Chopra and Paterson 1981; Karato 1986), lowering the viscosity by up to two orders of magnitude (Hirth and Kohlstedt 1996). In addition to this, they appear to be particularly important for the operation of plate tectonics. The absence of plate tectonics on Venus has been ascribed to the dryness of this planet relative to Earth (e.g. Nimmo and McKenzie 1998). One important effect of water is the reduction of the strength of the lithosphere. Regenauer-Lieb et al. (2001) produced numerical models of the initiation of subduction by sediment loading. Using a complex, laboratory experiment based rheology which depends on the water content, they found that the presence of water promotes the localization of stresses and the formation of a shear zone through the entire lithosphere. It is clear that the breaking of the lithosphere is an important first step for starting subduction.

Water also reduces the solidus and liquidus temperatures of mantle peridotite by up to several hundred Kelvin (e.g. Mysen and Boettcher 1975; Katz et al. 2003). As a result, a wet mantle undergoes more chemical differentiation through partial melting than a dry mantle. The generally basaltic melt products may find their way to the (near) surface to add material to the crust. The depleted residue which remains in the mantle has a lower density than the starting material (Jordan 1979), and may act to stabilize the mantle (Parmentier and Hess 1992; De Smet et al. 2000; Zaranek and Parmentier 2004). On the other hand, basaltic crustal material may, when reaching a sufficiently high pressure, transform into dense eclogite (Hacker et al. 2003), which can destabilize the crust and lithosphere (Dupeyrat and Sotin 1995; Van Thienen et al. 2004b). The balance between these two effects remains to be investigated in detail.

Combined, the effects of mantle weakening and increased melt production by volatiles result in an increased capacity for cooling of the mantle in the plate tectonics scenario. The former effect causes faster advection of heat in the mantle. The latter results in more advection of heat by melts to the (near) surface. For the stagnant lid scenario, the net effect is less clear, as increased heat transport may be mitigated by a decreased advective and conductive mantle heat flow due to a stabilizing, depleted buoyant layer in the upper part of the mantle (Schott et al. 2002; Zaranek and Parmentier 2004).

3.1.3 Regime Transitions

A relatively simple way to consider regime transitions is by considering the energetics of convection. Stevenson (2003b) wrote this energy balance down as follows for an active lid (plate tectonics) scenario:

$$\text{gravitational energy release} = \text{viscous dissipation of the internal flow} \\ + \text{work done deforming the plates.} \quad (3)$$

The second term of the right hand side takes place primarily at subduction zones (Conrad and Hager 2001). This term obviously depends on the mechanical properties of the lithosphere, which in turn depend on temperature and water content. From energetic considerations, one may predict a negative correlation between the magnitude of this term and the number of plates, with a single plate planet as an end member case (Stevenson 2003b).

Sleep (2000) studied the transitions between the three major modes of mantle convection, magma ocean, plate tectonics and stagnant lid, in the context of secular cooling of terrestrial planets, primarily considering the effects and degree of melting (see Fig. 2). Note that in this context, the term magma ocean refers to a partially molten region (mush) within the planetary interior, in contrast to an accretion-related initial magma ocean which extends to the surface. The following presents an overview of these melting-related and additional processes which may cause or be part of a regime transition.

When the mantle temperature falls below the level which allows partial melting to take place underneath the mid-ocean ridges, ridge lock occurs. In the absence of axial magma chambers and associated systems the ridges now become mechanically strong and no longer easily support spreading (Sleep 2000). In this case, plate tectonics stops and a stagnant lid begins.

A transition from plate tectonics to a magma ocean takes place when the amount of heat removed from the mantle by plate tectonics is less than the amount of heat produced in the mantle. The mantle heats up and at some point, the crust which is formed at MORs does not completely solidify to its base anymore (Sleep 2000), resulting in a trench lock as well. Note that thin lids on top of a magma ocean tend to founder, causing the magma ocean to extend to the surface. A sufficiently high crystal fraction in the mush may mitigate this (Abe 1997; Elkins-Tanton et al. 2005).

Alternatively, a locking at the trenches may also occur when oceanic lithosphere becomes insubductable due to its inherent low density. This results in a transition to the stagnant lid regime. Generally speaking, a thicker, less easily subductable lithosphere is created for higher mantle temperatures and/or a lower gravitational acceleration (Sleep and Windley 1982; Vlaar and Van den Berg 1991; Davies 1980). Results of 1D melting models by Van Thienen et al. (2004c) show that the upper limits of potential mantle temperature allowing plate tectonics are about 1500°C for Earth, 1450°C for Venus, and 1300–1400° for Mars, based on lithosphere buoyancy considerations at the subduction zone (which is likely an oversimplification, Conrad and Lithgow-Bertelloni 2002; Labrosse and Jaupart 2007).

Mechanical instability of the lithosphere in a stagnant lid regime may cause the initiation of plate tectonics (Sleep 2000). Small-scale sublithospheric convection may also initiate subduction (Solomatov 2004), possibly starting the plate tectonics cycle. However, the initiation of subduction remains poorly understood (Stern 2004).

When the amount of heat generated in a planet's interior is greater than the amount which can be removed by conduction through a stagnant lid, the mantle may heat up to a point where large-scale partial melting starts to take place (Reese et al. 1998). The crust which is produced from this melt can either start subducting, thus initiating plate tectonics, or remain stable, thus retaining a low surface heat flow and continuing the heating up of the interior, and result in a magma ocean (Sleep 2000). In general, big planets produce thinner, more easily subductable crust than small planets (Van Thienen et al. 2004c), so the former would tend to evolve to the plate tectonics regime whereas a magma ocean would be a more likely outcome in the latter case (Sleep 2000).

In the model of Sleep (2000), the magma ocean freezes by upward solidification of the mush up to the base of the lid. This causes a transition to plate tectonics if part of the lid can subduct or to a stagnant lid regime if not. Note, however, that differentiation during the solidification of a magma ocean results in an unstable density stratification which may overturn to produce a very stable density stratification (Elkins-Tanton et al. 2003, 2005), thus inhibiting the onset of convection in the mantle.

A mechanism for the cessation of a hypothetical early phase of plate tectonics on Mars (e.g. Sleep 1994) was proposed by Lenardic et al. (2004). They suggest that the thick southern hemisphere crust may have grown from a small original size to such an extent that it reduced the mantle heat flow to a point where the mantle stopped cooling down and started heating up again. As the mantle heated up, the viscosity dropped and convective stresses dropped below a critical value required to mobilize the lithosphere.

A geodynamical regime transition in the other direction would also be possible, when cooling of the planetary interior results in an increase in convective stresses. Lenardic et al. (2004) suggest on the basis of similar arguments that Earth may have had a stagnant lid phase before plate tectonics became active. Based on thermal calculations, one would expect Mars to have switched back to plate tectonics as well. Possible explanations why this has not happened include the high rigidity of the Martian lithosphere and chemical buoyancy of the thick crust stabilizing the lithosphere.

Many of the changes in geodynamical regime discussed above are related to melting, as discussed by Sleep (2000), and thus changes in the internal temperature of a planet. However, recent work by Loddoch et al. (2006) demonstrates that, at least for the stagnant lid case, this is not a necessity. Their 3D numerical convection models show incidental mobilization of a stagnant lid in thermally equilibrated models in an apparent statistical steady state.

3.2 Magnetic Field Generation, Requirements and Effects

The existence of a global magnetic field is important for the habitability on terrestrial planets as it can protect life from severe radiation. Without the magnetic field, life will have to retreat to places covered by enough rock to be shielded from damaging radiation effects, a circumstance that could limit biosphere diversification. In addition to the direct influence of the magnetic field on life, there exists an important cycle, which relates the dynamo generation to the stability of the atmosphere, the surface conditions and the interior dynamics. Before the description of these feedback mechanisms, we introduce the general concept of magnetic field generation and its requirements.

The global magnetic fields of terrestrial planets are generated in their iron-rich and liquid cores. Although the dynamo mechanism is imperfectly understood, it is generally accepted that core convection is needed, as well as a finite rotation rate of the body, though the slow rate of Venus already suffices (Stevenson 2003a). Thus, a necessary condition for a dynamo is the presence of convection, which can be either by thermal or compositional convection.

Thermal convection in the core, like thermal convection in the mantle, is driven by a sufficiently large super-adiabatic temperature difference between the core and the mantle. It occurs if the core heat flow supersedes that conducted along the core adiabat. The latter heat flow, therefore, serves as a criterion for the existence of thermally driven convection in the core. If the mantle removes heat from the core at a rate that exceeds the critical heat flow, then the core will convect. If the mantle removes heat at a rate less than the critical heat flow, the core is thermally stably stratified; dynamo action by thermal convection would not be possible. To cool the core above the threshold is difficult to achieve and it is suggested that a purely thermal dynamo may exist only for a short time of a planets evolution. The most likely period is just after accretion, in particular if the core is superheated with respect to the mantle. The initially strong cooling rates decline rapidly, possibly within a few thousands to a few hundred million years, to values where the heat flow out of the core can be accommodated by conduction alone. How long thermal convection can be sustained depends on the efficiency of mantle cooling. The more efficient the mantle cooling, the longer the existence of a thermal dynamo (e.g. Breuer and Spohn 2006). The existence of a thermal dynamo can be further prolonged in case of radioactive heating in the core. It is usually assumed that the terrestrial cores do not contain a significant amount of any radioactive elements, but it has been suggested that a larger concentration of potassium could have been incorporated at high pressure in the core during its formation (e.g. Lee and Jeanloz 2003).

Compositional density differences are 100% efficient in driving convection, in contrast to temperature differences (described by the Carnot efficiency factor, which is between 0.06–0.11, e.g. Braginsky 1964; Stevenson et al. 1983; Braginsky and Roberts 1995; Lister and Buffett 1995). The mechanical work provided by convection equals the viscous and ohmic dissipation (Hewitt et al. 1975; Backus 1975; Braginsky and Roberts 1995; Lister and Buffett 1995; Labrosse 2003), the former of which can be neglected in the core (Braginsky and Roberts 1995), and the latter of which is the expression of the magnetic field. Compositional convection can occur due to the release of positively buoyant light material during the process of solid inner core freezing from a fluid core with non-eutectic composition (Braginsky 1964). The dominant light element is suggested to be sulphur but also silicon, oxygen or hydrogen are possible candidates. For Earth, oxygen may be the most likely (Alfè et al. 2002).

Compositional convection and the associated generation of a magnetic field in the core occur if the temperature in the fluid (outer) core ranges between the solidus and the liquidus of the core material. As a consequence of the formation of an inner, nearly pure iron-rich core, the outer fluid becomes enriched in the lighter elements resulting in a decrease of the melting temperature. The melting temperature of the eutectic composition is in general very low (e.g. Fei et al. 1997) and it is very difficult to totally freeze the core of a terrestrial planet. This circumstance suggests that once a chemical dynamo has started—which is presumably late in the evolution—its existence is likely to be longstanding and it is difficult to stop as long as the core is cooling. Whether an inner core can grow depends on the initial amount of light elements in the core and on the efficiency to cool the interior below the liquidus. The more light elements in the core the lower is the melting temperature and the more difficult to cool the core below that temperature.

For Mercury it is suggested that the sulphur content is only a few percent and core freezing starts with respectively high core temperatures. A different situation is given for Mars, where a sulphur content of about 14% is suggested from SNC meteorites (Wänke and Dreibus 1994). Here, the melting temperature is comparatively low. In such a case, the interior needs to cool efficiently to initiate inner core formation and a chemical dynamo. Planets with one-plate tectonics cool inefficiently whereas planets with plate tectonics cool comparative efficiently. Thus, for the latter, inner core growth is much more likely, e.g. for the Earth where the inner core growth has initiated about 1 to 3.5 Ga ago (e.g. Labrosse et al. 2001; Gubbins et al. 2003). Note however, that the probability for a one-plate planet to cool below the core liquidus is enhanced for a wet and weak mantle rheology and/or for a thin mantle as it might be the case for Ganymede and Mercury, respectively.

3.3 Interior Atmosphere Interaction

Surface conditions such as atmospheric pressure, temperature and the presence of water are key parameters for planetary habitability. These may strongly depend on interactions between the outer and inner layers of the planets. Most of the publications concerning mantle dynamics and volatile exchange focus on Earth (e.g. McGovern and Schubert 1989; Franck and Bounama 1995, 1997, 2001) and few involve complete two-way coupling between interior and exterior (to some extent in Franck et al. 1999; Phillips et al. 2001a). Here we mainly focus on Mars and Venus since the interactions are simpler for these planets, but nevertheless illustrate the most important processes.

The planetary science data available mostly consist of atmospheric measurements and surface observations, since these are most easily obtained, and less of data pertaining to planetary interiors. The data suggest both planets are currently uninhabitable. They lack stable liquid water at the surface, and have either a tenuous atmosphere (Mars) or one that is much too hot and dense (Venus). Despite these observations, we can tell that in the past, things were certainly different.

3.3.1 Mars

Mars in particular shows remnants of water-related features such as outflow channels and valley networks which are evidence of the existence of liquid water (stable or not) on the surface at various times in Mars' past for unknown periods of time (Carr 1996; Jakosky and Phillips 2001; and Chaps. 3 and 4 of this volume). Moreover, missions such as Mars Express with its High Resolution Stereo Camera and the OMEGA spectrometer have provided us with new data on the morphology and composition of certain areas of Mars. Hydrated minerals have been found and also sulfate-rich layered deposits in Valles Marineris (Bibring et al. 2005), as well as in the north polar region (Squyres et al. 2004; Langevin et al. 2005). These sulfates imply that liquid water was present during the formation of the layers, though not necessarily as stable bodies of water. Other observations (e.g. ASPERA data) have allowed estimates for the present atmospheric escape rate of Mars, suggesting a value of about 1 kg/s (see Carlsson et al. 2006). This implies an evaporation of the present atmosphere in some 200 Myr at present conditions. The HRSC has revealed recent episodic volcanic activity on Mars (Neukum et al. 2004). Calderas on major volcanoes present signs of activation and resurfacing during the last 800 Myr with events that may be as young as 4 Myr. Recent detection of methane, with a photochemical lifetime of around 300 years, in Mars' atmosphere (Krasnopolsky et al. 2004; Formisano et al. 2004) suggests a presently active source. Both a biogenic origin (Formisano et al. 2004) and a geological origin (Lyons et al. 2004) have been proposed. Lyons et al. (2004) show that limited recent dike-like activity and hydrothermal alteration can account for the observations, which would support continued addition of volatile species to the Martian atmosphere from its interior. As discussed in Sect. 2.2, the polar layered depositits and ice caps are major agents in the present-day surface-atmosphere interactions of water and CO_2.

These observations suggest that in the past and in the present, volatiles have been and still are released into the atmosphere of Mars and have likely been essential to the evolution of the planet. The primary interaction between the atmosphere of a terrestrial planet and its mantle is indeed the degassing of volatiles (water and CO_2 in particular). It is commonly admitted that early hydrodynamic escape and heavy bombardment occurring during the first hundreds of million years of the planets history efficiently remove the primordial gaseous layer leaving up to 1% of the early CO_2-rich atmosphere (see Carr 1996). Thus volcanism

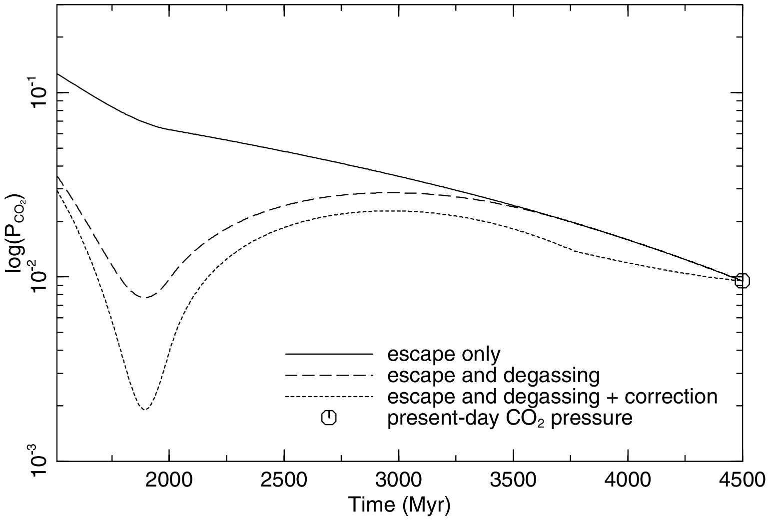

Fig. 3 Evolution of the maximum CO_2 pressure in the atmosphere of Mars over the last three billion years. The *solid curve* shows only atmospheric escape as described in Chassefière et al. (2007). The *dashed curve* shows escape and includes degassing of CO_2 by crust production using results from Breuer and Spohn (2006). The *dotted curve* uses the same model as the *dashed one* but adds a corrective factor for late activity which is not present in the crust production model

can have a strong influence over medium to late atmospheric history. It is however difficult to quantify the amount of degassing for two reasons. First, data is scarce and melt production rate estimates can be obtained either by observation of resurfacing (Hartmann and Neukum 2001) or with numerical modeling by looking at the crust production. Both methods lack the desired accuracy and provide only an order of magnitude estimate at best. The second difficulty originates from uncertainties in the composition of the planet's interior, which, for Mars, is rather poorly constrained. It is nonetheless clear that inner dynamics determine the amount of degassing by controlling partial melting through the temperature of the mantle and therefore influence the evolution of the atmosphere. A simple numerical model taking into account only degassing (from crust production data, see Breuer and Spohn 2006) and non-thermal atmospheric escape (Chassefière et al. 2007) can be used to provide an estimate of the history of CO_2 pressure on Mars that fits with present conditions (Fig. 3).

In these models, most, if not all, of the present day CO_2 has been released by volcanism, as the primordial atmosphere is blown away during the first hundreds of million years by bombardment and hydrodynamic escape. Only low pressures which are insufficient to create enough greenhouse effect to obtain a warm climate during the past three billion years are observed. This is consistent with the current consensus that a dense (>1 bar CO_2) atmosphere could not have persisted more than some tens of millions of years before collapsing, with CO_2 forming carbonates that no one has been able to detect so far (Catling and Leovy 2006).

Phillips et al. (2001b) calculated that the emplacement of Tharsis may have resulted in the degassing of $1.7 \cdot 10^{22}$ g water and $1.9 \cdot 10^{21}$ g carbon in the form of CO_2. Compared to the present-day atmosphere inventory (see Table 1), these are vary large quantities.

3.3.2 Venus

For Mars, most of the interaction between the interior and the atmosphere appears to take place in the form of mass and heat transport from the former towards the latter reservoir. However, in the opposite direction, the atmosphere or hydrosphere of a terrestrial planet may influence the dynamics of the solid interior, primarily through the action of water (see Sect. 3.1.2). In addition to this, surface conditions provide the upper thermal boundary condition for the solid planet. Thus, a dense and hot atmosphere can delay the cooling of the planetary mantle, affecting its thermal and also its magmatic/volcanic evolution. Phillips et al. (2001a) investigated such a coupled evolution for Venus. They used a simple parameterized mantle convection model (including partial melting) and a radiative-convective atmospheric model. Feedback was incorporated by release of water associated with melt extraction into the Venusian atmosphere which adds to the greenhouse effect and thus affects the surface temperature. After two billion years of evolution the coupled model was hotter by several tens of K and its extrusive volcanic flux was four times higher than the case without coupling. Their study shows a complex interplay between Venus convective evolution, volcanic activity and atmospheric state even with such a simple model featuring only basic processes. More complex and elaborate models are required to obtain a more detailed understanding of the interactions.

3.4 Life, CO_2 Fixation and Climate

CO_2 is a major greenhouse gas impacting habitability and Earth global climate. On geological time scales, the atmospheric CO_2 budget is mostly controlled by volcanic degassing, and consumption through the coupling of silicate weathering and carbonate precipitation (Berner et al. 1983), known as the Urey reaction (see reactions (4)–(6)):

$$(\mathrm{Ca,Mg,Fe})\mathrm{SiO_3} + 2\mathrm{H_2CO_3} = (\mathrm{Ca^{2+}, Mg^{2+}, Fe^{2+}}) + 2\mathrm{HCO_3^-} + \mathrm{SiO_2} + \mathrm{H_2O} \tag{4}$$

(weathering of silicates, mostly on continents)

$$\mathrm{Ca^{2+}, Mg^{2+}, Fe^{2+}} + 2\mathrm{HCO_3^-} = (\mathrm{Ca,Mg,Fe})\mathrm{CO_3} + \mathrm{H_2CO_3} \tag{5}$$

(cations and HCO_3^- are consumed by carbonate precipitation in oceans)

$$(\mathrm{Ca,Mg,Fe})\mathrm{SiO_3} + \mathrm{CO_2} = (\mathrm{Ca,Mg,Fe})\mathrm{CO_3} + \mathrm{SiO_2} \tag{6}$$

(net budget showing that C is transferred from the atmosphere to sediments through silicate weathering).

Silicate weathering and carbonate precipitation are thus two important processes to consider, and we discuss here the impact of life on them.

3.4.1 Diversity of Factors Affecting Silicate Weathering

Many studies have focused on river geochemistry, including small watersheds or at a larger scale using the main rivers worldwide to infer parameters governing weathering (for a review, see Dupre et al. 2003). Weathering rates can also be determined in laboratory experiments conducted over short periods of time or from analysis of soil profiles formed over geological timescales. Simple parametric laws describing the dependency of weathering on diverse environmental factors are then established and introduced into numerical models calculating the evolution of atmospheric CO_2 over geological times. The environmental factors impacting silicate weathering that are classically considered are: (1) climate (Chadwick et al. 2003), silicate weathering increasing with temperature (White et al. 1999) and runoff, both enhanced under high CO_2 values; (2) Lithology of the weathered rocks, basalt weathering acting in particular as a major atmospheric CO_2 sink (Dessert et al. 2003); (3) the age of the rock substrates (White and Brantley 2003); (4) the intensity of physical denudation with the important role of soils on chemical weathering (Gaillardet et al. 1999); (5) Configuration of continents, etc. … Life is an additional parameter, whose quantitative role on silicate weathering is however still poorly constrained.

3.4.2 Life as an Enhancer or Inhibitor of Silicate Weathering

There are numerous mechanisms mediated by life that may potentially enhance silicate weathering (e.g. Schwartzman and Volk 1991): e.g. release of CO_2 inducing a weak amplification of weathering rate (Von Bloh et al. 2003), lowering of local pH (Barker et al. 1997), trapping of nutrients freed by weathering, physical weathering through microfracturing of mineral grains, release of organic acids and ion-complexing organic ligands (organic molecules which may act as ion carriers) like siderophores (e.g. Liermann et al. 2000), modification of Fe redox modifying Fe-containing silicates dissolution (for a review, see Barker et al. 1997). The debate over which effects are most significant is important as some mechanisms may be a passive result for other processes, whereas others involve an active input of energy providing them a different status from an evolutionary perspective.

Several studies have evidenced a biological enhancement of silicate weathering either in laboratory experiments or in the field, at the watershed scale (e.g. Aghamiri and Schwartzman 2002; Moulton and Berner 1998; Lucas 2001; Hutchens et al. 2003; Rogers and Bennett 2004). However, several processes mediated by life could also inhibit silicate dissolution: e.g. production of organic polymers reducing silicate surface reactivity and/or the water/rock ratio, enhancement of secondary phase precipitation passivating silicate surfaces, and modification of iron redox, lowering iron-containing silicate reactivity (e.g. Santelli et al. 2001). Several studies have evidenced a biological inhibition of silicate weathering both in laboratory and field observations (e.g. Santelli et al. 2001; Ullman et al. 1996; Benzerara et al. 2004; 2005a). A study of Hawaiian basalt flows of known age indicates that there is an acceleration of chemical weathering on the order of at least 10 to 100 times for lichen covered surfaces relative to bare rock (Jackson and Keller 1970). Moulton and Berner (1998) indicated an enhancement factor of two to five in a Western Iceland catchment. Aghamiri and Schwartzman (2002) evidenced problems of elemental incongruency as enhancement factors were found to be 16 considering Mg^{2+} and only 4 considering Si.

Considering the high number of mechanisms potentially controlling silicate weathering, there is probably no single value for a biological enhancement factor, which might vary depending on biological communities and external physico-chemical conditions (Kelly et al. 1998). Hence, many more studies at the watershed scale should be systematically conducted to find potential constants in the biological enhancement/inhibition of silicate weathering.

3.4.3 Impact of Vascular Plants

An important question is whether vascular land plants have a higher effect on silicate weathering than microbes, as a significant difference between the impact of microbes and plants on silicate weathering would imply a qualitative and quantitative shift of the modalities of silicate weathering over Earth history as vascular plants emerged in the Silurian (440 Ma ago).

The point is not to present microbes and plantes as intrinsically opposite agents, as most of plants nowadays live in close association with microbes. Moulton and Berner (1998) concluded from their study that trees enhance silicate weathering more than mosses and lichen partially covering rocks. However, this kind of study should be conducted more systematically and pertinent comparisons should be made whenever possible.

3.4.4 Impact of Life on Carbonate Precipitation

The precipitation of carbonates (reaction (5)) is an additional step of the Urey reaction where life can have a dramatic impact. There is no unequivocal evidence that calcium carbonate can precipitate abiotically from seawater (e.g. Wright and Oren 2005).

Although calcium carbonate and dolomite should precipitate spontaneously from supersaturated solutions in seawater based on thermodynamic considerations, they do not (Lippmann 1973), and it has been suggested that kinetic inhibitors to carbonate precipitation are effective. On the other hand, there is increasing evidence that these inhibitors can be overcome through microbial activity by e.g. modifications of solution chemistry, control of pH at the microscale, removal of kinetic inhibitors (for a review, see Wright and Oren 2005). In marine environments throughout the Phanerozoic, carbonate precipitation has been considered almost exclusively a biological process mediated by skeleton-forming invertebrates, algae or protozoa. Could the formation of Precambrian carbonates have been mediated by microbiological processes? Or might they have been simply chemical precipitates? Would carbonates precipitate in the absence of life? This is one of the major questions in the present discussion and remains as a key issue in geobiology (e.g. Arp et al. 2001; Wright and Oren 2005).

Numerous experimental studies have shown that microorganisms can promote precipitation of carbonates (e.g. Castanier 1999; Morita 1980), but their importance in the natural environment at the global scale has yet to be quantified. Different mechanisms can be involved: e.g. removal of CO_2 from the medium by autotrophic pathways (e.g. Riding 2000), active ionic exchanges of Ca^{2+}/Mg^{2+} through the cell membrane, production of carbonate ions and pH increase by ammonification, dissimilatory reduction of nitrate or degradation of urea, production of carbonates, pH increase and removal of sulphates which are inhibitors for calcium carbonate precipitation by dissimilatory reduction of sulphates (e.g. Vasconcelos et al. 1995). Microbes can also provide preferential nucleation sites. It is still much debated whether the polymers they excrete are enhancers or inhibitors of calcium carbonate precipitation (e.g. Arp et al. 1999; DeFarge et al. 1996). Recent observations show that those polymers are widespread in many biological carbonate deposits and suggest that they do have an impact on carbonate precipitation (e.g. Gautret et al. 2004; Benzerara et al. 2006).

3.4.5 Microbial Urey Reaction at a Much Shorter Time Scale?

The Urey reaction is usually seen as a two-step process on a time scale of hundreds of thousands of years, with silicate weathering occurring on continents and carbonates precipitating in oceans (Berner 1995). It is interesting to note however that few studies have shown the possibility that the two steps occur in a much shorter time and at the same location when catalyzed by microbes (e.g. Ferris et al. 1994; Rogers and Bennett 2004; Benzerara et al. 2005b). The importance of such a microbial promoted Urey reaction needs to be evaluated.

3.4.6 Modeling of the Quantitative Effect of Life on Climate Regulation

The Urey reaction can provide a purely abiotic negative feedback on atmospheric CO_2 content regulation. Indeed, an increase of atmospheric CO_2 raises global temperatures which enhance silicate weathering and hence CO_2 trapping. If life does significantly enhance silicate weathering, then the efficiency of this self-regulating process is greatly amplified and becomes more sensitive to any perturbation. This has been considered as the best established Gaian negative feedback: if global temperatures increase, biological growth increases and rock weathering increases, which ultimately reduces carbon dioxide in the atmosphere. As temperature decreases due to the reduction in atmospheric CO_2 (a greenhouse gas), life-mediated silicate weathering decreases. It is noticed that it is a non-selective feedback on growth: if any studies have suggested that biological weathering strategies have evolved because they provide a source of limiting nutrients like phosphorus or iron (e.g. Lenton 1998; Bennett 2001; Rogers and Bennett 2004), they were not selected for their effect on global carbon dioxide content. Changes in atmospheric carbon dioxide content or temperature do likely not alter the selective advantage of a strategy for nutrient search. Hence, this feedback does not drive Earth to a state intrinsically dependent on the presence of life and responding only passively to external forcing.

Several studies have incorporated the biological amplification of the Urey reaction in models of the global carbon cycle over geological time scales (e.g. Lenton 1998; Lenton and Lovelock 2001; Von Bloh et al. 2003). A stronger biotic amplification of weathering implies a stronger negative feedback. It moreover provides a responsive regulation against perturbations that tend to increase carbon dioxide and/or temperature (Von Bloh et al. 2003). Von Bloh et al. (2003) show that specific values of the biotic amplification on silicate weathering extend the life span of the biosphere. Refined models taking in account other processes as well as the measure of the biotic effect on silicate weathering throughout Earth history would improve our knowledge of the stability of the Earth system.

4 Planetary Evolution

As we have treated the processes relevant for life and habitability in the previous sections, we will now move on to discuss these in the context of the (geodynamical) evolution of planets Earth, Mars and Venus. Whereas Earth appears to have had clement surface conditions for several billion years, there is evidence for significant changes in the conditions at the Martian and Venusian surfaces during their evolution. Along the same vein, Mars and Venus show signs of possible changes in the geodynamical regime, the former very early in its history and the latter relatively recently. Earth shows evidence for the continued operation of plate tectonics over a long period of time, starting 1 up to 4 billion years ago (De Wit 1998; Hamilton 1998; Stern 2005). Changes in the thermal evolution of these planets are implied by the changes in the dynamical evolution.

4.1 Thermal Evolution Time Scales

The thermal evolution of a planet is controlled by the heat balance equation,

$$MC\frac{\partial(T-T_s)}{\partial t} = -Q + H(t), \tag{7}$$

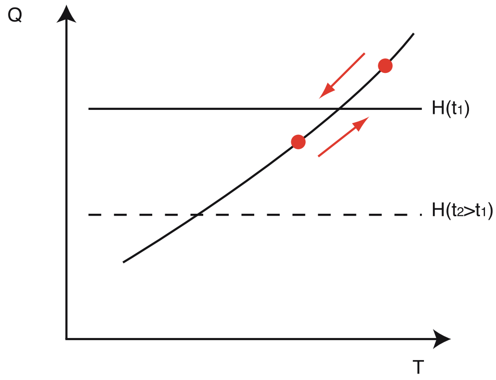

Fig. 4 Schematic thermal evolution of a planet. The direction of the evolution depends on the balance between the heating term and the heat flow term in (7). Scaling of mantle convection provides a relationship between the heat flow and the potential temperature of the planet. The monotonic decay of radiogenic heating eventually leads to cooling of the planets

with M the mass of the planet, C its averaged heat capacity, T the potential temperature, T_s the surface temperature, Q the heat flow at the surface and H the time decaying radiogenic heat production. Several isotopes contribute to $H(t)$ and each has a separate time-scale, ranging from less than 1 Gyr to 20 Gyr. From an initial temperature T_0, the planet can either cool or heat up, depending on the sign of the right-hand side of (7), but the decay of heat production will eventually make this sign negative and lead the planet to cooling (Fig. 4). The time scale with which the temperature decay follows the decay of heat production is given by comparing the left-hand side of (7) and the first term on the right-hand side:

$$\tau = \frac{MC(T - T_s)}{Q}. \tag{8}$$

The heat flow can be measured at the surface of the planet (it is rather well known at the surface of the present Earth, as will be discussed below) but is subject to important evolution with time. The first order effect comes from the evolution of temperature itself. Indeed, the heat flow per unit surface of a planet can be written as $q = k(T - T_s)/\delta$, with k the thermal conductivity, and δ the thickness of the boundary layer at the top of the mantle. When the interior of the planet cools, the heat flow decreases, linearly if the boundary layer thickness is kept constant. This is however usually not the case, since this parameter is the result of dynamical balances in the mantle, in which for example the temperature dependent viscosity plays an important role.

Most models of the thermal evolution of planets use scaling laws that relate the heat flow at the surface to the potential temperature (e.g. Turcotte and Oxburgh 1967; Schubert et al. 1980; Christensen 1985; Solomatov 1995; Reese et al. 1998). These scaling laws are obtained using numerical or analog models of thermal convection and simple dynamical arguments that balance the dominant stresses in the system. The difficulty in this exercise comes from the identification of the dominant balance in the considered planet which, as discussed above (Sect. 3.1) and citing Sleep (2000), can evolve on different regimes, each having a different $Q(T)$ scaling (Fig. 2). Some of these scalings are rather well established, like for example the stagnant lid convection (e.g. Davaille and Jaupart 1993; Solomatov 1995) but the conditions for transition between regimes are still poorly understood (see Sect. 3.1.3). Sleep (2000) emphasizes the role of melting which is clearly a key unresolved issue in the problem. Melting of course depends on temperature but also on composition, particularly in volatile elements, and influences volatile extraction and rheology. This means that using the potential temperature as unique parameter controlling the heat flow is probably not correct.

One effect that has not been sufficiently investigated yet and might be important in the context of habitability is the evolution of the surface temperature T_s which is controlled by the radiative balance of the atmosphere (see Sect. 3.3). The amount and nature of the chemical species present in the atmosphere depend in part on the degassing history of the internal part. Small variations of T_s have a tremendous effect on life but very large variations and disparities among planets can also occur, as exemplified by the difference between the Earth and Venus.

4.2 Thermal and Dynamical Evolution of Earth

The Earth is the best place to start constructing models of thermal evolution since the amount of available data places strong constraints on the general procedures developed in these models. The history of this topic shows that the issue is far from obvious and the points of view of geochemistry have traditionally been orthogonal to that of geophysical modeling, due to the difficulty in satisfying constraints concerning both the present heat production in the mantle and the heat flow at the surface. The reason for this difficulty can be understood by getting a closer look at the way most geophysical models work.

The thermal evolution of the Earth is modeled using (7) and usually a scaling relationship of the form

$$Q = AT^{1+\beta}\nu^{-\beta} \tag{9}$$

with A a scaling constant, ν the temperature dependent viscosity of the mantle and β an exponent close to $1/3$ in usual boundary layer theory (e.g. Schubert 1979). The temperature dependence of the viscosity is responsible for a short adjustment time scale τ. Following Christensen (1985), one can assume a viscosity law of the form $\nu = \nu_0(T/T_0)^{-n}$ and linearly develop (7) around T_0 to obtain as adjustment time scale:

$$\tau_a = \frac{MCT_0}{(1+\beta+\beta n)Q_0}. \tag{10}$$

Using the classical value $\beta = 1/3$ and $n \simeq 35$ that approximate best the Arrhenius law of mantle materials (Davies 1980; Christensen 1985) gives $\tau_a \sim 800$ Ma. This leads to a heat flow at the mantle surface that closely follows the decay of heat production, and at least 70% of the present surface heat flow is radiogenic (the Urey number is 0.7). On the other hand, the geochemical models of the Earth favor a present radiogenic heat production of about 20 TW or less (Javoy 1999) to be compared to the present total heat loss of 46 TW (see Jaupart et al. 2007, for a comprehensive discussion of the energetics of Earth's mantle).

A classical way to solve the issue is to change the exponent β to a lower value, as was first done by Christensen (1985). Indeed, (10) shows that the thermal adjustment time scale is then increased and the heat flow decay with time can lag much longer after the decay of radiogenic heating. Several models have been proposed to lead to a smaller β: effect of temperature dependent viscosity (Christensen 1985), resistance of plate to bending (Conrad and Hager 1999), chemical buoyancy of a thicker crust associated with deeper melting at high temperature (Sleep 2000; Korenaga et al. 2003). All these effects may well be important but no evidence has been shown to support the Earth being in the regime where they actually are.

Another view of this problem has been recently proposed by Labrosse and Jaupart (2007) who use the observed distribution of surface heat flux as input to a thermal evolution model, therefore not relying on any scaling assumption. Because of the difficulties associated with the measurements of seafloor heat flux, the heat loss of the Earth is usually computed using the distribution of seafloor ages, that is linked to the heat flux using a half-space cooling model that give

$$q = \frac{kT}{\sqrt{\pi \kappa a}}, \tag{11}$$

where a is the age of the seafloor and κ is the thermal diffusivity. Everywhere a good sediment coverage allows reliable heat flux measurement, the data agrees very well with this model (Jaupart et al. 2007).

Using maps of seafloor ages (Sclater et al. 1980; Müller et al. 1997) one can obtain a distribution of seafloor ages (Sclater et al. 1980; Rowley 2002; Cogné and Humler 2004) which can be used to obtain the total heat loss through the oceans in the form

$$Q_{\mathrm{oc}} = \lambda(f) \frac{A_{\mathrm{oc}} kT}{\sqrt{\pi \kappa a_{\max}}}, \tag{12}$$

$\lambda(f)$ being a geometrical factor depending on the distribution of seafloor ages, presently equal to $8/3$, $A_{\mathrm{oc}} = 3.09 \cdot 10^{14}$ m^2 the total oceanic surface and $a_{\max} = 180$ Ma the maximum age of the oceanic lithosphere.

The continental contribution to the heat loss of the Earth is more difficult to measure because the continental heat flux varies greatly, particularly on short length scales. These variations cannot be attributed simply to geologic ages but essentially reflect the variability in concentration in radioactive elements of crustal rocks. In fact, the heat flux at the surface can be attributed in a large part to heat production in the crust and the flux from the mantle has been estimated to values of the order of 7–15 mW m^{-2} (Pinet et al. 1991; Jaupart et al. 2007), to be compared to the average of 100 mW m^{-2} of the oceans. It is then reasonable to assume the continents to be perfect insulators when modeling the thermal evolution of the Earth (Grigné and Labrosse 2001). Equation (12) can then be used in the heat balance (7), after removing the heat flow at the surface of continents from the radiogenic heat production term.

Assuming that $\lambda(f)$, A_{oc} and $a_{\max}$ have been constant through time, the heat balance (7) can be solved (Labrosse and Jaupart 2007) and the resulting evolution of temperature and heat flow are shown in Fig. 5. The main outcome of this model is that the heat flow from the seafloor has been essentially constant through time and the total change of temperature is of order 150 K. Such a mild thermal evolution could have been anticipated by estimating the Earth's thermal adjustment time scale according to (8) which, using (12) gives

$$\tau_E = \frac{MC\sqrt{\pi \kappa a_{\max}}}{A_{\mathrm{oc}} k \lambda(f)} = 10 \text{ Gyr}. \tag{13}$$

Another interesting feature of this model is that equilibrium between radiogenic heat production and heat loss is reached about 3 Gyr ago. Whether the warming of the Earth for times prior to that actually occurred depends on the validity of the assumptions made. In particular, the total oceanic surface may not have been constant before that (Collerson and Kamber 1999).

The assumption that could be subject to most criticism is that of $a_{\max}$ being constant. Another assumption that could be made is that its value is set by the stability of the lithosphere, that is to say that the Rayleigh number computed with the lithosphere thickness at that age,

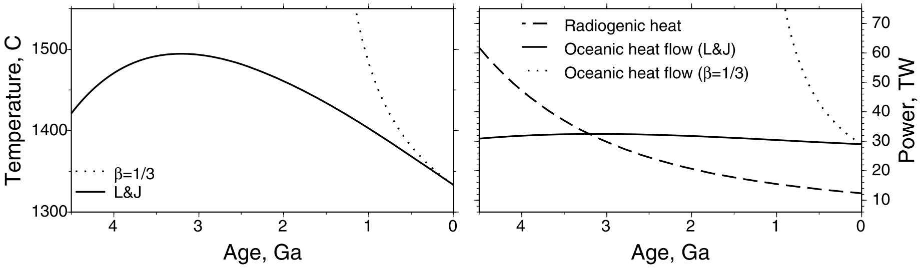

Fig. 5 Thermal evolution of the Earth based on the observed distribution of seafloor ages (*solid lines*, labeled L&J) or from a boundary layer stability hypothesis (*dotted lines*, labeled $\beta = 1/3$). After Labrosse and Jaupart (2007)

$\sqrt{\pi\kappa a_{\max}}$, is equal to the critical one R_c, that is

$$\sqrt{\pi\kappa a_{\max}} = \left(\frac{R_c\kappa\nu}{\alpha g T}\right)^{1/3}. \tag{14}$$

With this assumption we recover the classical scaling given by (9) with $\beta = 1/3$ and the numerical solution to the heat balance equation is shown as dotted lines in Fig. 5. This shows that the assumption that the maximum age of the oceanic lithosphere is set by its stability cannot be accepted. The distribution of seafloor ages actually shows that the Earth is able to subduct a lithosphere of any age with a more or less equal probability (Rowley 2002) and this justifies our assumption of a constant maximum age. Alternatively, the assumption that the mechanism resulting in (11) and (12) has been active since Earth's early history may be wrong. Geologists are still divided on when plate tectonics started to operate (see e.g. Stern 2005), and the other terrestrial planets in our solar system warn that plate tectonics should not be taken for granted.

On the other hand, the geometrical coefficient $\lambda(f)$ clearly depends on the distribution of continents at the surface of the Earth and has a large effect on the heat flow at the surface. If the distribution of seafloor ages was flat, the value of λ would be 2, instead of the present $8/3$, which would give a 30% decrease of the oceanic heat flow compared to the present value. This means that large fluctuations of the heat flow on the time scale of 500 Myr are to be expected and their magnitude overwhelms its long term evolution (Grigné et al. 2005).

4.3 Thermal and Dynamical Evolution of Mars and Venus

The surface morphologies of Mars and Venus differ significantly from that of Earth most likely mirroring fundamental differences in the present-day tectonic regimes and heat-loss mechanisms of these planets. The Earth has a dichotomy between oceanic and continental regions and features mid-oceanic ridges, orogenic mountain belts and island arc systems, features that are commonly linked to plate tectonics. No significant plate tectonic features have been identified on Mars and Venus although on the former the linearity of the Tharsis volcano chain has been likened to island arc volcanic chains on Earth (Sleep 1994). More recently, crustal remnant magnetization patterns have been used to suggest some form of plate tectonics for early Mars (see Connerney et al. 2004 for a recent review and Connerney et al. 2005 for most recent data). Early plate tectonics have also been proposed (Nimmo and Stevenson 2000; Breuer and Spohn 2003) to allow for a better cooling of the core thereby facilitating the explanation of an early magnetic field. For Venus, early plate tectonics have been suggested (e.g. Phillips and Hansen 1998) but as for Mars there is little, if any, factual evidence for it. It is commonly held that plate tectonics requires fluid water on the surface to operate, the water modifying the rheology of the cold plates such that subduction becomes feasible. The high D/H ratio in the Venusian atmosphere strongly suggests but does not prove that it had substantial water in the past (e.g. Fegley Jr. 2004). Today Venus is desiccated as far as the available spacecraft data can tell.

4.3.1 Volcanism

Mars has a major volcanic center, Tharsis, that covers almost a hemisphere and rises up to 20 km above the datum. There is a second much smaller center, Elysium. Other volcanoes are distributed more randomly across the surface. The surface of the planet shows a dichotomy between roughly the southern hemisphere and the northern hemisphere. While the southern hemisphere is old (about 4 to 4.5 Ga) the northern hemisphere is about 1 Ga younger (Zuber 2001). Whether or not the northern hemisphere has been resurfaced by volcanic activity or by sedimentation is debated. Mars Global Surveyor gravity data show (Smith et al. 1999) that an old cratered surface exists buried underneath the northern hemisphere. Tharsis is centered almost at the border between these two hemispheres. Volcanic activity on Tharsis has apparently continued until the very recent past, albeit at a small rate (Neukum et al. 2004). The recent knowledge on Mars volcanism has been summarized by Solomon et al. (2005).

Venus has no distinct volcanic center but a prominent rift system, the Beta, Atla and Themis Regiones. Volcanic features include a large number of comparatively small shield volcanoes that are apparently randomly distributed, volcanic plains and about 500 coronae—circular large scale volcano-tectonic features. It is widely accepted that coronae form above upwelling plumes (Smrekar and Stofan 1999), though downwellings rather than upwellings have been suggested as well (Hoogenboom and Houseman 2006). Johnson and Richards (2003) have recently reanalyzed the coronae distribution and have concluded that the coronae activity at least tended to concentrate in the Beta, Atla and Themis rift system over time. The paucity of impact craters on the surface has been used to suggest that Venus was resurfaced about 500–700 million years ago by a volcanic event of global scale (Schaber et al. 1992; Bullock et al. 1993; McKinnon et al. 1997). This event was followed, as the model suggests, by almost no volcanic activity. The suggestion prompted a number of papers that speculated on an episodic thermal history of Venus with pulses of global activity followed by epochs of relative tranquillity like the present one (e.g. Turcotte 1995; Phillips and Hansen 1998; Reese et al. 1999; Ogawa 2000; Nimmo 2002). The early conclusions from the cratering record have been challenged by e.g. Hauck et al. (1998) and Campbell (1999) and most recently by Bond and Warner (2006). According to the latter authors, the cratering record allows a variety of interpretations in terms of volcanic resurfacing including a global decrease in time in the rate of an otherwise statistically distributed volcanic activity. Previous workers had still concluded that Venus underwent a major transition in tectonic style, albeit more gradual than the previously postulated sudden global resurfacing event. The surface geology also seems to indicate that modifications of the surface are planet wide and and gradual over long periods of time rather than episodic and locally confined (Ivanov and Head 2006).

Whether or not there is present day volcanic activity on Venus is unclear. Some sulphuric clouds have been interpreted as being evidence of recent volcanism (Fegley and Prinn 1989). The recent data of the infrared spectrometer Virtis on Venus Express (e.g. Helbert and

Benkhoffm 2006) suggest differences in surface temperature that can be linked to morphologic features on the surface but it is too early for more far reaching conclusions.

4.3.2 Thermal and Dynamical Evolution

Since planets can be regarded as heat engines that convert thermal energy to mechanical work, part of which is necessary to maintain the magnetic field against losses by ohmic dissipation, the evolution of the planet subsequent to formation and early differentiation is strongly coupled to its thermal evolution or cooling history. Also, as noted above, the cooling history is strongly related to the history of the magnetic field. In the context of the discussion of the evolution of habitability of a planet and even life, the outgassing history matters. The latter is again linked to the thermal history of the planet (e.g. McGovern and Schubert 1989; Phillips et al. 2001a). Phillips et al. (2001a) have presented models where the evolutions of the Venusian atmosphere greenhouse and the interior are coupled. These models suggest that the coupling can be quite effective for Venus.

Early models of the thermal histories of Mars and Venus (e.g. Schubert 1979; Stevenson et al. 1983; Schubert and Spohn 1990; Spohn 1990) used simple parameterized convection schemes that did not include effects of the temperature dependence of the viscosity on the convection. But Turcotte (1995) had already noted the difficulty for Venus to remove its internal heat if it lacked plate tectonics as the surface geology suggests. Following the work of Davaille and Jaupart (1993), Solomatov (1995), Moresi and Solomatov (1995) and Grasset and Parmentier (1998), stagnant lid models of planetary convection and thermal history were developed and successfully applied to Mars (e.g. Hauck and Phillips 2002; Breuer and Spohn 2003: Breuer and Spohn 2006; see also Spohn et al. 2001 and Connerney et al. 2004 for recent reviews). These models suggest that Mars is continuously cooling at a rate of about 30 K/Ga. Volcanic activity and thereby outgassing decreased rapidly during the first few 100 million years and gradually from there on to non-zero but almost negligible (volume wise) values during the past Ga. On Mars, outgassing would therefore have contributed to or generated an atmosphere early on that was subsequently lost due to atmospheric loss processes in the absence of the protecting magnetic field. The thermal history calculations would principally allow for a wet early Mars. But the geologic record as revealed by the HRSC camera on Mars Express and the MER suggests a desert planet that had wet spots with surface water in places of volcanic activity that melted the permafrost (e.g. Head et al. 2005).

Although the continuously cooling thermal history models of Mars are successful in explaining the observations (e.g. crustal production, magnetic field), it is not excluded that the Martian thermal evolution could be episodic. The Rayleigh number for Martian mantle convection is comparatively small (10^5 to 10^6), a fact that allows for quasi stationary convection but the interaction of the flow with possible mantle phase transitions such as the olivine to spinel and the spinel to perovskite phase transitions may introduce time dependence. For instance, since the spinel-perovskite transition is strongly temperature dependent an early perovskite layer at the bottom of the lower mantle of Mars may thin and eventually disappear (Spohn et al. 1998) and allow for increased heat flow from the core. However, presently available data do not constrain the present-day presence or absence of the perovskite layer (Van Thienen et al. 2006). The interaction of Martian mantle convection with phase transitions has been used before to explain the singularity as of Tharsis as volcanic center (Harder and Christensen 1996; Breuer et al. 1996, 1997, 1998) but most recently Schumacher and Breuer (2006) have criticized the hypothesis of Tharsis being maintained by a mantle superplume and have introduced a model of thermal blanketing and mantle melting, instead.

The thermal history of Venus appears to be more complicated (see Schubert et al. 1997, for a review). Although it must in principle be possible to remove the heat generated by radiogenic elements in a stagnant lid convection regime the resulting thermal lithosphere may be too thin to be consistent with the available gravity data (Anderson and Smrekar 2006). The latter data suggest an elastic lithosphere of up to 100 km thickness and a crust of 90 km thickness. The necessity of removing the heat in the absence of plate tectonics leads to episodic scenarios as have been proposed by Turcotte (1993), Turcotte (1995), Turcotte et al. (1999), Reese et al. (1999), and Ogawa (2000). Some authors (Turcotte 1995; Phillips and Hansen 1998; Nimmo 2002; and most recently Van Thienen et al. 2005) suggest that present Venus may actually be heating up. This would offer another explanation of the absence of a dynamo magnetic field on the planet. Stevenson et al. (1983) had attributed the absence of a field to the absence of an inner core and a cooling rate of the core that was too small for convection and dynamo action to occur. If instead Venus had an inner core but a mantle the temperature of which would be increasing in time, then the inner core would have ceased to grow (it may actually be shrinking in radius) and the dynamo could not be driven by chemical buoyancy released upon inner core growth. Chemical buoyancy is believed to be the main driver of the Earths dynamo (e.g. Labrosse 2003; Stevenson 2003a).

The D/H ratio on Venus has been used to infer that Venus had a wet past with water on the surface with the equivalent of a global ocean of up to about 500 m depth (Donahue and Russell 1997). It is not known when the transition from a more habitable climate to the present desiccated state occurred, but see Grinspoon and Bullock (2005) for a recent discussion of the water history of the planet. It can be speculated that the removal of water from the atmosphere was accompanied by a transition from plate tectonics to the present tectonic state. The transition may have been accompanied by the loss of the magnetic field. Whether or not remnant crustal magnetization may have survived is speculative. If observed by future space missions, it would suggest a more habitable past of the planet.

5 Discussion and Important Open Questions

5.1 Discussion

Looking at Tables 1 and 2, it is clear that the amounts of volatiles present in Earth's interior are greater by far than the amounts available to the biosphere. The fluxes of the listed species are such that the entire atmospheric inventory is processed through the mantle on time scales significantly shorter than the age of planet: on the order of 10^6 yr (C), 10^8 yr (S), and 10^9 yr (H_2O), respectively (Note, however, that the errors in these estimates are large due to the large uncertainties in the fluxes). Therefore, the internal dynamics have the potential to significantly affect the availability of these species to the biosphere. Indeed, the balances listed in Table 2 show that the internal dynamics are constantly changing the reservoir sizes. Nevertheless, life's involvement in the fixation of volatile species, specifically C and N, allowing them to be subducted eventually, is also important; not so much for the mantle reservoir, but certainly for the hydrosphere and atmosphere reservoirs. In addition to this, life may contribute to determining the amount which is subducted. More specific, in the case of sulfur the oxidation state of the ocean, depending on the availability of oxygen, determines the amount of sulfur which is deposited in the form of pyrite in oceanic sediments, which may eventually be subducted. For carbon, the total volcanic flux is only three times as large as the subduction flux. This means that the biosphere, through its capability to precipitate carbonates, has the capacity to noticeably influence the amount of carbon (see Sect. 3.4) available in the hydrosphere/atmosphere on epoch time scales (this is apart from the fixation of carbon in organic matter on much shorter time scales). It is hard to quantify this due to uncertainties in the numbers of Tables 1 and 2, variations of these values with time, and also the partial decoupling of carbonate deposition in the ocean basins and subduction of these carbonates (see Sect. 2.4.3).

Because of the size of the solid planet, these biological processes do not significantly influence the composition of the mantle. However, through their regulation of the atmospheric CO_2 concentration, they strongly influence the greenhouse effect and thus surface conditions in terms of temperature and availability of water. As we have shown in Sect. 4.1, the surface temperature affects the rate of convection and cooling of the mantle. The availability of water may determine whether the lithosphere is weak enough to break (Sect. 3.1) and allow subduction zones, an essential ingredient of plate tectonics, to be formed.

Such considerations also illustrate the importance of plate tectonics compared to one-plate convection. As illustrated in Fig. 1, the latter regime is generally thought not to allow the return of volatiles into the planet's interior. As a consequence, it is more difficult for any fixated volatile to be made available again to the biosphere. When plate tectonics case is active, elevated temperatures in the mantle contribute to releasing at least a part of the fixated material again in volatile form. However, some return of volatiles into the mantle may be possible, even in the stagnant lid case, by delamination of lower crustal material (e.g. Dupeyrat and Sotin 1995; Van Thienen et al. 2004a, 2004b; Elkins-Tanton 2007a, 2007b), assuming the sinking material contains volatile species. On Venus, the high atmospheric pressure may prevent exsolution of volatile species from lavas at the surface. In this case, return of the volatiles into the atmosphere is also possible by devolatilization or partial melting of sinking material.

Plate tectonics with volcanism at ridges and subduction zones is a continuous regime both in time and space, whereas the stagnant lid regime with plume and local tectonics-related volcanism is not. This is a fundamental difference. Earth's mid-ocean ridges form a single connected grid (Ricou 2004). At these ridges, production of new crust is more or less continuous. A recent compilation of sea floor ages suggests more or less constant spreading rates over the past 180 Myr (Rowley 2002). Subduction related volcanism is neither spatially nor temporally continuous, but the discontinuities both in space and time are limited. Reorganization, creation, and destruction of ridges and subduction zones take place as part of the Wilson cycle (see Turcotte and Schubert 2002). However, this does not interfere with the continuous existence of spreading centers and subduction zones.

On the other hand, volcanism related to plume activity (whether caused by a plume or an alternative mechanism, see e.g. King and Anderson 1995) is by definition limited to individual isolated localities. Also from a temporal point of view, activity is limited, starting with the eruption of flood basalts, and ending with the termination of the plume tail. Geological evidence for an increase in size and frequency of flood basalt eruptions for the younger Earth has been presented by Abbott and Isley (2002), corresponding to a decrease in distance in both space and time. However, Ernst et al. (2005) note that there are still many uncertainties which hinder such trend analyses.

As volcanism provides a source of nutrient elements, and contributes to the cycling of volatile species which play a role in the atmospheric energy balance (see Sect. 3.3), the distinction between temporal continuity and episodicity corresponds to a distinction between: (a) continuous and episodic supply of nutrient elements and (b) a stabilizing versus a destabilizing effect on planetary climate. The extent of the first effect also depends on the characteristic residence time of the nutrient elements in the biosphere before burial in a geological reservoir. The latter effect is illustrated by modeling studies. For example, Phillips et al.

(2001b) showed that the injection of CO_2 and H_2O into the Martian atmosphere by volcanism associated with the formation of Tharsis resulted in a warmer climate, which declined afterwards.

Episodic clement climate conditions and nutrient availability require a form of life which is capable of surviving intermediate periods, which may be very long (on the order of millions of years or more). Moreover, it may require some mobility in case the availability of nutrients is limited to the location of volcanic activity and these locations change with time.

On the other hand, flood volcanism events may act to significantly disrupt an already established biosphere by affecting climate. The eruption of flood basalts coincides with several large mass extinctions (e.g. Courtillot et al. 1986; Renne et al. 1995; White 2002; Courtillot 2003). Though the largest mass extinctions may have required the combined effects of flood volcanism and a bolide impact (White and Saunders 2005), it is clear that massive volcanic eruptions may cause a cooling of the climate on short time scales through silicate and sulphate aerosols (McCormick et al. 1995) and warming on a longer time scale through the injection of massive amounts of carbon into the atmosphere as described in Sect. 2.4.2 which results in a moderate increase of atmospheric CO_2 levels (Berner 2002).

As mentioned above, the existence of a dynamo can be important for the habitability on terrestrial planets due to the feedback mechanisms between the interior and the atmosphere (Fig. 1): The magnetic field protects the atmosphere from erosion of light elements, primarily hydrogen, due to sputtering and/or hydrodynamic effects (see Chap. 6). A sufficiently thick atmosphere with the associated greenhouse effect is necessary to allow fluid water to be present on the surface. However, water will not be stable in a liquid state if the atmosphere is too thick and the greenhouse effect is too strong. Water affects the evolution of a planetary mantle and the planetary tectonic engine. First, it makes the lithosphere deformable enough for subduction of the crust to occur. Second, water recycled to the mantle by plate tectonics reduces both the activation energy for creep and the solidus temperature of mantle rock, thereby enhancing the cooling of the interior. At the end of the cycle, the sufficient cooling by plate tectonics and/or a wet mantle rheology may allow the formation of an inner core and thus, a longstanding magnetic field. In the long term, an interruption or a change of this cycle may result in conditions that are not favorable for life. For example, the cessation of plate tectonics results in the heating of the interior. As a consequence the core cannot cool, the inner core growth stops and the magnetic field vanishes. This starts a runaway process through which atmosphere erosion and loss is promoted; the surface conditions change and water may become unstable at the surface. A similar scenario could be observed in case of one-plate tectonics throughout the evolution. Due to inefficient cooling of the interior, inner core growth never starts. An early thermal dynamo, which exists only in case of a superheated core as a consequence of core formation, vanishes rapidly after few hundred million years. Thus, an early atmosphere can be eroded efficiently due to the lack of a global protecting magnetic field. Whether and how long the surface conditions allow the existence of water during the planets evolution may depend strongly on the density and composition of the early atmosphere.

5.2 Important Open Questions

An important question which needs to be investigated is to what extent life is essential for the long-term stability of water at a planet's surface. Would the Earth's atmosphere be hot like Venus' atmosphere, boiling away all the water, if life would not have arisen in the Archean and started fixating CO_2 from the atmosphere? Would the absence of oceans in such a scenario have prevented the operation (or even onset) of plate tectonics on Earth?

Would this, and the higher surface temperature, have reduced the heat flow from the mantle to such an extent that the core dynamo could not be sustained and Earth's protection against the solar wind would be removed?

At the moment we can only speculate on the answers to these questions, but they may hold the key to understanding why the Earth and its sister Venus and brother Mars are so alike in terms of composition and size, yet so different in terms of habitability.

Although it was primarily defined by astrophysicists, habitability is a central notion to which geobiologists and even geodynamicists can contribute. Life itself may modify habitability. This idea was central to the Gaia theory (e.g. Lovelock and Watson 1982). Although this theory has been largely criticized there is potentially an important role of life in the global carbon dioxide cycle at geological time scales. A better understanding of the mechanisms involved in this process, as well as a much better quantitative view of the biological role have yet to be obtained. Ironically, while some effort is done to look for extraterrestrial life, we need for our purpose to envision the evolution of a purely abiotic Earth which seems so far a remote aim.

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