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Strange relatives – Earth and her sibling neighbours*

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Abstract

Comparative planetology has brought to light some general themes in the geological evolution of planets and planetary bodies. Particularly the study of the Moon and the Earth-like planets Venus and Mars has contributed much to the understanding of the early evolution of the Earth. All have sizes and distances to the Sun of the same order of magnitude, are primarily composed of cosmically rare silicates and metals, and underwent differentiation into core, mantle and crust. The crust is basaltic on Venus and Mars, while on the Moon the crust is anorthositic in the highlands and basaltic in the maria. Earth has, besides a basaltic crust in the ocean basins, as the only planet in the Solar System a continental crust of average dioritic composition. This continental crust owes its genesis to the interaction between ocean water, basaltic crust, and mantle dynamics. All planetary bodies suffered an intense bombardment of meteoroids early in their histories. Impacting was probably the dominant geological process in that time, but on Earth virtually no traces of the early impacts are preserved because of later geological activities. The differences in environmental conditions and geological evolution between Venus, Mars and Earth are primarily determined by differences in size (cooling rate and gravity) and distance to the Sun (solar energy input). Only Earth has a hospitable environment suitable for life – conditions that are maintained by the biosphere. On Venus infernal conditions prevail, with a surface temperature of 480° C, an atmosphere almost entirely composed of CO_2, an atmospheric surface pressure of 88 bar, and no water. Very little is known about tectonic activities because of the dense sulfuric-acid cloud cover, but volcanism appears to be still active. Mars is a barren desert with extreme variations in temperature, and a very thin atmosphere with a surface pressure of 0.0064 bar almost entirely composed of CO_2. Water is present as ice in polar caps and in permafrost. Both Venus and Mars possessed early in their histories copious amounts of water (and Mars possibly also life ?). Venus lost all water because of 'runaway greenhouse' conditions, while Mars probably lost most of his volatiles because of the early episode of large-scale impacting – which ravaged all planets, but was more effective on the martian atmosphere because of the planet's smaller gravity. There are no indications that plate tectonics was ever in operation on Mars, but until fairly recently (and maybe up to now) there was active volcanism, giving rise to the largest volcanoes in the Solar System.

Beyond the Earth

Until some 25 years ago, Geology was the science of the Earth, while the other planets in our Solar System were the domain of Astronomy. However, planetary exploration over the last two decades has

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changed this situation. The planets and planetary bodies of our Solar System are nowadays objects of geological rather than astronomical study. Our approach to the geological exploration of the planets has been the reverse of that used over the last two centuries in the exploration of the Earth. From space we *begin* with a global view, and only later move to detailed observations and measurements in selected regions. This approach has proven to be a powerful one, but the rapid progress made to date was possible only because geoscientists are able to extrapolate from their knowledge of geological processes learned on Earth. Comparative planetology has brought to light some general themes in the geological evolution of planets and planetary bodies. These themes have, in turn, added to our perception of geological activities in the first 700 million years or so of Earth's history, the pre-Archaean of *Hadean* (Cloud, 1988), a period from which no geological record has been preserved.

Briefly, the solar System consists of nine planets, some 60 satellites and innumerable smaller bodies, all orbiting the Sun. The planets and their satellites can be divided into two broad classes, the Jovian or Jupiter-like bodies, and the terrestrial or Earth-like bodies. The Jovian planets consist of rocky cores mantled by hydrogen and helium and minor amounts of methane, ammonia and water. The terrestrial bodies include the planets Mercury, Venus, Earth and Mars (the inner planets), the Moon, and a number of satellites of the Jovian planets. They consist primarily of rock and metal with no gaseous compounds or only a small amount of gas in a thin atmospheric layer. The inner planets are roughly similar in size, mass, composition and distance to the Sun. Nevertheless, despite many similarities, they do vary widely in both geological and environmental conditions. Only Earth has the combination of temperature, atmosphere and abundant liquid H_2O necessary to sustain life. By contrast, the natural environments of Mercury, Venus and Mars are hostile to life.

Earth's perspective

Earth is an 'oasis in space' (Cloud, 1988), a perfect planet from a human point of view and the only habitable planet we know of. However, from a planetary point of view our planet is an improbable entity, an oddity among her sister planets. The most conspicuous anomalies are the abundance of liquid H_2O, the exotic chemistry of the atmosphere, the presence of the biosphere with an almost infinite variety of living creatures, and the topographic dichotomy of the surface. The latter reflects the fundamental compositional difference between the continental crust and the crust underlying the ocean basins: the oceanic crust being of rather uniform basaltic composition, and the heterogeneous continental crust having the *average* composition of diorite (Taylor & McLennan, 1985). No other planet shows a similar dichotomy. From a planetary point of view it is the terrestrial continental crust which is the oddity, as basalt is a universal rock, widespread on the planetary objects of our Solar System, whereas terrestrial-type continental crust is absent or very rare outside Earth.

All these anomalous features of our planet – the oceans, the atmosphere, the biosphere and the continents – are interrelated. They form the component parts of a complex, dynamic, interacting system of feedback mechanisms, which is unique to the Earth. It is this system, sustained by the energy flux from the Sun and the interior heat production of the Earth, that has governed Earth's evolution ever since very early in her history.

Let us first address the similarities and differences between our Earth and the neighbouring worlds of Venus and Mars. Venus in particular was once considered to be our twin sister by virtue of her similarity in size to the Earth. Until the recent advance in planetary exploration it was therefore naively expected that there should be a rather close correspondence between the conditions prevailing on both planets. However, as we presently know, Venus differs in many respects from the Earth. She has a searingly high surface temperature, and a waterless environment with an atmosphere nearly ninety times as dense as on Earth – a surface pressure equal to that at a depth of 900 metres below sea level.

Dry hothouse and glaciated desert

The copiousness of water is, at first sight, probably one of the most striking anomalies of our planet. Most of the water is in the oceans and seas, which cover some 70% of the Earth's surface. The wealth of water is why photographs taken from space show our planet a sapphire-blue globe capped by brilliant white fields of polar ice and flecked with clouds, in sharp contrast to the drab uniformity of our neighbours Mars and Venus, both lacking that abundance of water.

Venus is a dry planet, virtually devoid of water. Still, Venus must once have had a moist environment. Evidence for this is provided by the isotopic analysis of the trace amounts of water vapour in Venus' atmosphere by the Pioneer spacecraft. They show that venusian water has a D/H ratio of about 10^{-2} (Donahue et al., 1982; McElroy et al., 1982), versus about $1.5 \cdot 10^{-4}$ on Earth, in meteorites and (as was measured by the Giotto spacecraft, Eberhardt et al., 1987) in the comet Halley – an enrichment by a factor of 50–100 in the heavy hydrogen isotope deuterium. This is explained by isotope fractionation during the processes that led to the loss of substantial quantities of water due to the prevailing temperatures far above the boiling point; the present surface temperature is about 480° C, sufficient to melt tin and lead. Under such conditions water vapour was effectively transported to the upper atmosphere, where the water molecules split by ultraviolet light to form hydrogen atoms that escaped into space (Donahue et al., 1982; Kasting, 1989; Kasting et al., 1984, 1988). In the course of this process, the hydrogen that escaped was enriched in the lighter isotope, whereas the residual water in the atmosphere became enriched in deuterium. The 'left behind' oxygen was stirred back to the hot venusian crust and bound, mainly by the conversion of FeO to Fe_2O_3.

Whether the initial water was present in liquid form, or from the very beginning solely as steam in the hot atmosphere, is unclear. Possible topographical expressions of the former presence of liquid H_2O, for example in the form of fossil channels such as do occur on Mars, lie still hidden beneath the permanent cover of clouds consisting of droplets of concentrated sulfuric acid and elemental sulfur (Saunders, 1987). The unveiling of such features, if they exist, has to wait until the high-resolution radar images of the venusian landscape by the American Magellan spacecraft become available. It has been argued (Kasting et al., 1984) that the available evidence is most easily explained if oceans of hot water were initially present. The surface temperature on young Venus would have been in the order of 80–100° C, allowing the existence of substantial amounts of liquid H_2O, possibly enough to provide an ocean in the order of a kilometre deep over the whole planet. Early in her history the present inferno took hold of the planet, however, and has been maintained ever since through the greenhouse effect of the massive atmosphere almost entirely consisting of CO_2 (Table 1), with a present surface pressure of 88 bar. The planetary environment became a *runaway greenhouse* and within a couple of hundreds of million years Venus boiled off all her water.

It is not Venus, but Mars that has a more Earth-like surface environment. Still, from a human point of view Mars is a hopelessly barren, cold desert. The air temperatures at the surface vary strongly from one region to another, with extreme diurnal fluctuations. The coldest area is the south pole, where the temperature always remains below the freezing point of CO_2, but at the equator the air temperature at the surface may rise to not far below the freezing point of water in the strong afternoon sun. The ground itself may then occasionally even be heated to above 0° C. Current conditions on Mars, notably the low temperature and the very thin atmosphere with a surface pressure of only 0.0064 bar (corresponding to the Earth's atmo-

Table 1. Atmospheres of the terrestrial planets[1]

	Venus	Earth	Mars
CO_2	96.5	0.035	95.3%
N_2	3.5	78.1	2.7%
O_2	traces	20.9	0.13%
Ar	0.007	0.93	1.6%
total pressure	88	1	0.0064 bar

[1] After Wayne (1985).

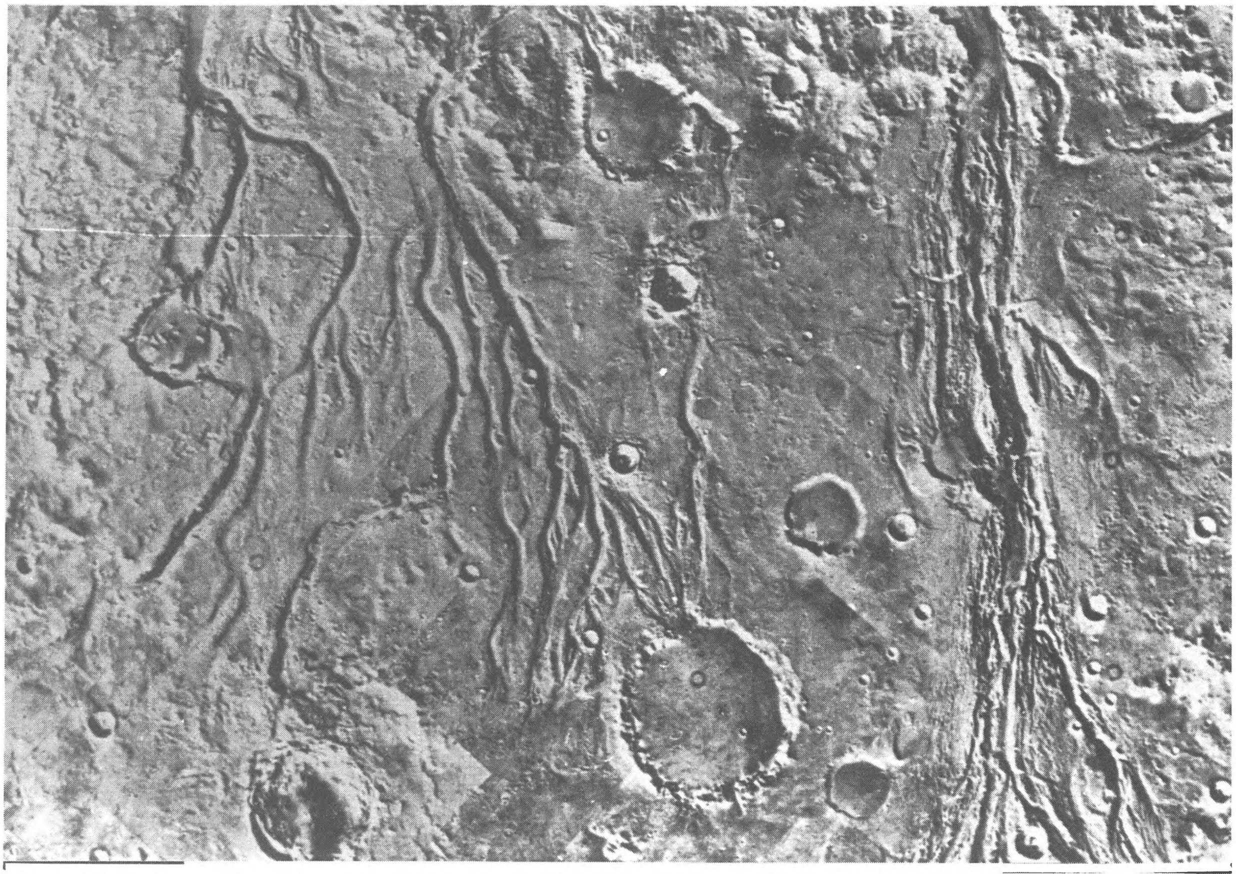

Fig. 1. Viking photomosaic of dendritic channel systems (runoff channels) on Mars. The valleys are deeply incised into an old, cratered terrain, draining Lunae Planum and flowing eastward into Chryse Planitia. North is to the left. Width equals about 200 km. Centered 17° N, 50° W.

sphere at an altitude of 40 km), do not permit the existence of liquid H_2O. Water does occur on Mars, but only in minor amounts in polar ice caps and as permafrost ice in the soil – the martian soil is permanently frozen to a depth of several hundred metres (Carr, 1987). Nevertheless, photographs taken by the Mariner and Viking spacecraft reveal channels which were clearly carved by running water, so copious amounts of water must once have flowed over the martian surface. These channels are loosely grouped in two categories (Sharp & Malin, 1975), the older *runoff channels* (Fig. 1) with tributary systems very much resembling dry river valleys (wadis) in terrestrial desert areas, and the larger and slightly younger *outflow channels* (Fig. 2), tremendous valley systems up to one thousand kilometres long and apparently formed by brief catastrophic floods due to the sudden release of huge reservoirs of underground water. A variety of mechanisms has been suggested to trigger such large floods (see, e.g., the review by Squyres, 1984), for example melting of permafrost ice by volcanic activity or as a consequence of large impacts, but probably several mechanisms are involved. The density of the superimposed impact craters (see below) puts the age of all fluvial features back to the remote past – the formation of the runoff channels being limited to the earliest part of the recorded geologic history, up to about 4 billion years ago, and the catastrophic floods taking place between approximately 4.0 and 3.5 billion years ago (Neukum & Hiller, 1981).

Mars has largely lost his initial water, along with most of the other atmospheric gases. This loss is

Fig. 2. Viking photomosaic of a catastrophic flood feature (outflow channel) on Mars. The 20 km wide channel arises from a region of chaotic terrain and flows eastward (to the right) into Simud Vallis. Centered 1° S, 42° E.

evidenced, for example, by the enhanced proportion of the heavy isotope ^{15}N in the martian atmosphere that was measured by the Viking spacecraft – a $^{15}N/^{14}N$ ratio of $5.94 \cdot 10^{-3}$ versus $3.66 \cdot 10^{-3}$ on Earth and in meteorites, an enrichment of about 62% (Biemann et al., 1976; McElroy et al., 1976). This provides a clue similar to the deuterium enrichment on Venus: a loss of atmospheric nitrogen to space with preferential escape of ^{14}N relative to ^{15}N. The Viking measurements also provide a clue to the time at which the loss of martian nitrogen took place: the abundance pattern of primordial noble gases (excluding isotopes that are produced from the decay of radionuclides in the planet's interior) equals that of the Earth's atmosphere, but the absolute abundances are nearly 2 orders of magnitude lower, while the $^{40}Ar/^{36}Ar$ ratio is $2{,}750 \pm 500$, about ten times higher than on Earth (Owen & Biemann, 1976; McElroy & Prather, 1981). This pattern is most easily explained by a massive loss of argon from the martian atmosphere *prior* to the buildup of ^{40}Ar through the decay of radioactive ^{40}K in the planetary interior, when ^{36}Ar was still more abundant than ^{40}Ar. It is inferred from these and other compositional properties of the atmosphere that Mars has outgassed and subsequently lost to space and to permafrost early in his history a quantity of water enought to provide an ocean at least some 10 to 100 m deep over the whole planet (Squyres, 1984; Pepin, 1986). Of the large quantity of CO_2 that was outgassed along with the H_2O, a minor part resides as dry ice in the polar caps and probably in the regolith, but the greater part is thought to be stored in carbonates which precipitated from ponds of water early in martian history (Kahn, 1985; Warren, 1987). The proof that crustal carbonates do occur has to await further exploration of Mars, however.

Our home planet: a hospitable anomaly

Earth has escaped both the catastrophic venusian heating and dehydration, and the martian atmospheric depletion and eternal glaciation. Our planet managed to establish early in her history overall surface temperature conditions within the range of liquid H_2O, as is evidenced by the nature of the oldest sedimentary rocks on Earth: the *ca* 3.8 billion year old deposits of the Isua supracrustal belt in western Greenland (Moorbath, 1977). The geologic record shows that since then Earth has steered a relatively constant temperature course, notwithstanding the steadily increasing energy output of the sun over the course of the Earth's history – every model of stellar evolution indicates that in the beginning of our Solar System, 4.55 billion years ago, the output of heat from the sun was about 30% less than it is now. This constancy is largely achieved by a climate-control system that operates through greenhouses gases in the atmosphere, particularly CO_2, and took hold of the Earth very early in our planet's evolution (Lovelock, 1982, 1989; Kasting et al., 1984).

This brings me to another conspicuous planetary anomaly of the Earth, the composition of the atmosphere. From a geochemical point of view, CO_2 ought to be the main constituent of the Earth's atmosphere, as this gas and H_2O are the principal volatiles released from the planet's interior through volcanic activities and, in the very beginning, through 'impact degassing' during the accretion stage (Matsui & Abe, 1986). Both Venus and Mars do indeed have atmospheres constituted for over 95% of CO_2 (Table 1), but on Earth the atmospheric CO_2 concentration amounts to only some 0.03%. Instead, the terrestrial atmosphere is made of a strange and incompatible mixture of gases that, from a chemical point of view, appears to violate the rules of chemical equilibrium (Lovelock, 1982, 1989; Wayne, 1985). In particular the concentrations of O_2 and N_2, and the trace amounts of reactive gases such as CH_4, N_2O, NH_3 and H_2 are far in excess of an equilibrium situation (Table 2). This unstable condition is maintained by interaction with the biosphere and lithosphere. Oxygen gas, for example, is produced by living plants through photosynthesis. The O_2 concentration in the atmosphere is kept more or less constant by a balance between the net gain of oxygen as carbon is buried in sediments, and the net loss by the oxidation of reduced materials in the crust. The excess carbon is locked up in the biosphere and as coal, oil, gas and dispersed carbon in sedimentary rocks. If the CO_2 inventory tied up in these reservoirs, in the oceans, and in the enormous deposits of limestone and dolomite, were to be added to the atmosphere, the partial pressure of CO_2 gas on Earth is estimated to rise to about 70 bar (Henderson-Sellers, 1983). This would be 200,000 times greater than the present pressure and of the same magnitude as in the venusian atmosphere, where a CO_2 pressure of 87 bar prevails. Because of the greenhouse effect, the terrestrial surface temperature would then be about 225° C – higher than even the worst doomsday predictions of global warming about which our politicians nowadays display so much concern. If all that CO_2 were back in the Earth's atmosphere, the proportion of N_2 gas relative to CO_2 would be similar to that on Venus and Mars, about 2%. The high concentration of O_2 gas in the atmosphere – in fact a waste product of the biosphere and as such an early case of global air pollution – is a prerequisite for the existence of multicellular life. Free oxygen also creates and maintains the ozone layer that protects life outside the marine environment against the lethal ultraviolet radiation of the sun. It

Table 2. Actual and equilibrium concentration (in %) of reactive gases in the Earth's atmosphere.[1]

	Expected fractional equilibrium concentration[2]	Actual fractional concentration
N_2	$<10^{-8}$	78.1
CH_4	$<10^{-33}$	$1.7.10^{-4}$
N_2O	$<10^{-18}$	$3.0.10^{-5}$
NH_3	$<10^{-33}$	10^{-7}
H_2	$<10^{-33}$	$5.3.10^{-5}$

[1] After Lovelock (1982); Wayne (1985).
[2] Values based on the current O_2 concentration of 21%.

is through the high O_2 concentration in the atmosphere that land life has become possible.

The terrestrial atmosphere has to be seen as an extension of the biosphere and hydrosphere. The biosphere also plays a dominant role in maintaining the surface temperature on Earth within the narrow range of liquid H_2O. It does so by regulating the concentration of the greenhouse gases in the atmosphere through a series of elaborate feedback loops to counter forces of change that may arise on our planet (Lovelock, 1982, 1989). If, for some reason, the atmospheric CO_2 concentration were to fall far below the present level, the Earth would cool to a frozen state. But then, biological activity and chemical weathering processes would slow down, leading to less removal of CO_2 from the atmosphere through the formation of biomass and carbonates, whereas the supply through volcanic gases and crustal decarbonation reactions would remain the same. The balance between supply and removal of CO_2 would then be shifted towards the supply side. This would cause the atmospheric CO_2 concentration to rise again, thus bringing about a rise in temperature. Conversely, if the CO_2 concentration were to rise above the present level, the resulting warmer, and consequently wetter climate would lead to an acceleration in biological activity and chemical weathering. This would make the balance between CO_2 supply and removal to shift towards the removal side. The CO_2 buildup and consequent rise in surface temperature would then be stemmed.

In short, if conditions become hotter, life flourishes and CO_2 is consumed. If conditions become colder, biological activities decline and CO_2 builds up again. It is through such a feedback loop that the biosphere has maintained through geologic time, by controlling the level of greenhouse gases in the atmosphere (besides CO_2, also CH_4 and NH_3), a nearly constant surface temperature that allowed the oceans to persist – which, in turn, is a prerequisite for the existence of life. The planetary network of intricate feedbacks that makes up this biological and geochemical control system and is responsible for maintaining the fine level of stability (homeostasis) on the Earth's environment, is nowadays widely referred to as *Gaia* (Lovelock, 1982, 1989). However, in the very beginning of Earth's history, before the development of the planetary ecosystem, the planetary environment must necessarily already have been in an 'abiological steady state' with temperature conditions in the range of liquid H_2O – if not, life would not have developed at all and no biologic climate-control system could have got a grip on our planet. How these conditions came into being, is still subject to much speculation. Probably, it resulted from the (fortuitous) combination of 'right' distance to the Sun and 'right' rate of geochemical recycling of CO_2 between atmosphere and crust (Kasting et al., 1988): the removal of CO_2 from the initial atmosphere by inorganic carbonate production was apparently faster than its addition by volcanism and other degassing processes, leading to a CO_2 level that, in combination with the dimmer Sun, kept or brought the temperature at the Earth's surface within the range of liquid H_2O. The rate of carbonate production must have been much higher in the distant past, as pristine rocks were exposed at a higher rate because of the more vigorous mantle convection regime (Priem, 1987) – since both the internal radiogenic heat production and the contribution of primordial heat sources (accretion, core separation) were much higher than today – and the large-scale brecciation of the crust by the high flux of impacting meteoroids (see below). Venus, although nearly equal in size and with the same initial atmospheric composition, failed to reach the same temperature conditions primarily because of the closer distance to the Sun: in combination with the thermal blanket provided by the atmospheric CO_2, this raised the surface temperature to above the boiling point of water and caused the present runaway greenhouse conditions. If, for some reason, the biologic control system that rules the Earth for at least the last 4 billion years were to fail, leading to global surface temperatures either above the boiling point or below the freezing point, all water on Earth would become either steam or ice. Life would then be terminated. It is hard to imagine how a planet ever could regain her comfortable climate once it has come in the venusian state of runaway greenhouse conditions, or once all water is frozen. Because of the high albedo, a sort of *runaway glaciation* would then set in.

Biosphere, atmosphere, climate and oceans are thus intimately linked in a single planetary, complex, self-regulating system. Earth is the only planet we know of where this system came into being. The planetary exclusiveness of the terrestrial continental crust appears also to be related to this system. There still is considerable debate about continental crust formation, but I take the view (Priem, 1987, 1988) that the continental crust has grown through geologic time and that the mechanisms leading to its formation can only proceed through the interaction between the oceans, basaltic crust and mantle dynamics. New basaltic crust is continuously generated from the mantle in spreading ridges, and reacts with ocean water during upwelling and during its journey to the subduction zone. This makes that the ocean-floor basalts, moving down into the mantle along subduction zones, differ chemically from the basaltic magma upwelling in the ridges: they have undergone hydration and other modifications. This alteration of mantle-derived basalts is a prerequisite for the generation in the subduction zone environment of the calc-alkaline (dioritic-andesitic) magmas that build through time the continental crust. If ocean-floor basalts would not have reacted with ocean water, the product of magma generation in the subducting oceanic plate ought to be basaltic again, with at most very minor amounts of calc-alkaline differentiates – if any. In short: *no oceans – no calc-alkaline magmatism – no continental crust.*

Moon: a battered world

The absence of oceans may thus be one of the factors why continental crust is absent or rare on other planets. But what do we know about the geology of our neighbours? So far, the best studied celestial body outside Earth is the Moon (see, e.g., the reviews by Taylor, 1975, and Ryder, 1987). Large quantities of lunar rock samples have been returned to Earth and were thoroughly studied. The basic geologic principles used to interpret lunar history are essentially the same as those used to study the history of terrestrial events: the law of superposition and the law of cross-cutting relations. Another method of determining the relative ages of lunar features is *crater chronology,* based on crater-frequency distributions – the longer a surface has been exposed, the more craters it will display (Taylor, 1975). A relative dating of different rock units can thus be obtained by counting the number of impact craters per unit surface. The relative order of events obtained by these means is provided with benchmarks of absolute time by the isotopic dating of returned samples.

The oldest parts of the lunar crust are the heavily cratered *lunar highlands,* mainly consisting of brecciated *anorthosites* with ages between 4.44 and about 4.0 billion years. The oldest age (Sm-Nd) of 4.44 ± 0.02 Ga (Lugmair, 1987) is believed to approach the crystallization age. The anorthositic crust is thought to have developed as cumulates of plagioclase feldspar by flotation in a completely dry basalt-magma ocean (Warren, 1985; Taylor, 1989), shortly after the formation of the Moon. A younger feature of the Moon are the sparsely cratered *lunar maria,* large basins more than 300 km in diameter which have been filled with repeated outpourings of *basaltic lava.* The basins were blasted out in the beginning of lunar history by huge impacting meteoroids. The volcanism in the maria started over 4.2 billion years ago (Taylor et al., 1983). It came essentially to an end about 3.2 billion years ago, but a few basalt flows may be of younger age (Ryder, 1987). The lunar eruptions did not create volcanic mountains, but took place along long fissures; lunar maria basalts are thus comparable to, for example, the lava plains of the Deccan Traps in India. All lunar rocks are much like the equivalent rock types on Earth, but they are severely depleted in volatile elements with regard to terrestrial rocks. Moon rocks are extremely dry and hydrated minerals like amphiboles, micas, etc., are conspicuously absent.

Most or all of the lunar volcanic activity ceased a little more than 3 billion years ago. Since then, very little has occurred. Any change at the surface is an extremely slow process – a million years from now, the footprints of Apollo astronauts in the lunar soil will still be fresh. Almost no global tectonic activity appears to have taken place. This means that the Moon 3 billion years ago looked much as it does today, contrary to the Earth, which would be unrecognizable even a few hundred million years ago. Because of the preservation of the early geologic record of the Moon, we know nowadays that *impact cratering* must have played an important, maybe a dominant role in early planetary history – also on Earth, where this time-span of unimaginable violence is appropriately named the Hadean. The combination of lunar crater counts, dated lunar samples, and counting of meteoroids in near-Earth space today, has shown that the lunar impact rate was about thousand times greater by 4.0 billion years ago than by 3.8 billion years ago and ever since (Taylor, 1975). Earlier, it may have been even higher, but most evidence concerning the lunar history before about 4.0 billion years ago has been obliterated by the heavy cratering. It is therefore still under debate whether the intense cratering around 4 billion years ago represents a unique event, such as the disruption of a large asteroid that scattered fragments throughout the inner Solar System, or the final stages of the condensation of the Solar System, when the leftover debris from the solar nebula were still crashing in on the already formed planets. On Earth, no evidence for any impact craters dating from this period of heavy bombardment has been preserved because of later vigorous geologic activity. However, on the basis of the highland record on the Moon, making due allowance for the differences in size and gravitation, it is estimated (Grieve, 1987; Grieve & Parmentier, 1985) that in Hadean time at least some 3,000 impact basins with a diameter greater than 100 km – of them maybe some 200 greater than 1,000 km – along with countless smaller craters, were formed on the surface of the Earth.

The bulk density of the Moon is 3.344 (the uncompressed density is 3.3, Taylor, 1988), only slightly more than that of the crustal rocks. There is thus little possibility for a significant increase in density with depth. The Apollo experiments included direct measurement of the rate of heat flow from the interior, while seismometers measured the response of the Moon to meteoritic impacts and moonquakes. The heat flow measurements point to higher temperatures at depth, due to radiogenic heat production and possibly a small remnant of the initial accretional and core-separation heat. From the seismicity it is clear that the Moon is layered, with a crust ranging from 60 to 100 km in thickness, a rigid mantle extending to a depth of 800 km, and probably a central core of partially molten rock (Taylor, 1975). Most moonquakes originate in the region near the base of the mantle. It is questionable whether the Moon has a metallic core; in any case, it has no magnetic field.

Earth's exotic twins

On Venus, the analyses by the Soviet Venera and Vega landers, and surface characteristics indicate that the crust is dominated by basaltic lavas (Basilevsky & Head, 1988). Gamma-ray data from the highland regions show that here also rocks occur high in potassium (4%), uranium (2.2 ppm) and thorium (6.5 ppm), which led to early speculations that granite is present on the surface of Venus, implying an evolution resembling that of the Earth. However, analyses by later landers make it more probable that we are dealing with alkali basalts or syenites (Taylor, 1989). The surface of Venus is relatively young, based on crater chronology on maps produced by radar imaging, but estimates vary widely from about one billion years (Ivanov et al., 1986) to as young as about 45 million years (Schaber et al., 1987). It is highly probable that there still is active volcanism on Venus. The evidence includes the much too high SO_2 concentration in the atmosphere, about ten times that which could co-exist in equilibrium with likely surface materials, and attributed to introduction into the atmosphere by geologically very young processes (Prinn, 1985); the large fluctuations in the concentrations of SO_2 in the atmosphere measured by orbiting spacecraft over the last 20 years, which are usually associated with outbursts of intense volcanic activity (Esposito, 1984); the landforms of apparently volcanic nature revealed by imaging radar (Prinn, 1985; Barsukov et al., 1986; Head & Wilson, 1986); and possibly the low-frequency radio emissions that seem to cluster around volcano-like landforms and have been interpreted as lightning discharges similar to those observed in the dust plumes of erupting volcanoes on Earth (Singh & Russell, 1986), although this interpretation is being contested (Taylor & Cloutier, 1986; Wood & Francis, 1989).

We have no direct information about the structure of the deep interior of Venus, since no seismometers have been deployed so far. The bulk density of 5.245 (uncompressed density of 4.0, Taylor, 1988) and the differentiation of a basaltic crust make rather certain, however, that Venus likewise is a layered planet, with a dense, metal-rich core, although the planet has no magnetic field. Whether plate-tectonic processes are (or were) at work on Venus, is still an unresolved issue. There are topographical features which resemble tectonic structures on Earth (Philips & Malin, 1984; Basilevsky & Head, 1988), and some authors (Head, 1990) even interpret these structures as indicating that processes of large-scale convergence, underthrusting and crustal imbrication continue to operate at the present time. We must await the results of the high-resolution radar exploration of Venus' surface by the Magellan spacecraft to look further into this question. The venusian atmosphere does provide us with some evidence, however, that the planet has been for a long time geologically less active than the Earth. We find this evidence in the isotopic and abundance data of atmospheric argon obtained by the Pioneer spacecraft: a $^{40}Ar/^{36}Ar$ ratio of 1.03 ± 0.04 (in comparison to 295.5 in Earth's and $2,750 \pm 500$ in Mars' atmosphere) and a total amount of ^{36}Ar about 70 times greater than that on Earth (Hoffman et al., 1980). Taking into account that Venus' mass is 0.81 that of the Earth, this implies that the total amount of released radiogenic ^{40}Ar is only about one fourth of the terrestrial value. This indicates that there has been much less outgassing from the mantle on our sister planet than on Earth, suggesting billions of years of less endogenic activity (Turcotte & Schubert, 1988). The volcanism on Venus is probably related to hot plumes rising through the mantle, and not to plate-tectonic processes (Prinn, 1985).

Mars is a much smaller planet, with a mass only 11% of that of the Earth. The total surface area is about equal to that of the continents on our planet. An outstanding feature of the surface of Mars is the mysterious north-south asymmetry (Arvidson et al., 1978). The northern hemisphere consists of rather smooth, sparsely cratered volcanic plains and huge volcanoes. The southern hemisphere, on the other hand, is a densely cratered highland; crater counting points to an age of over 4 billion years. A well-defined escarpment separates the two provinces. It has been suggested (Wilhelm & Squyres, 1984) that this twofold division is the result of a giant impact – the plains would largely fall within an ancient impact basin.

The American Viking landers analyzed surface material at two sites 4000 km apart in the southern hemisphere. The analysed samples consist of loose, fine material, analogous to loess on Earth, which appears to represent a dust layer covering most of Mars' surface and has the chemical composition of basalt (Toulmin et al., 1977). The volcanic plains of the northern hemisphere are younger and have a lower elevation than the southern highland. The volcanoes emerging from these northern plains (Carr, 1976) resemble in shape terrestrial shield volcanoes, like Hawaii. They have calderas at the summit, like most shield volcanoes on Earth. The largest volcano is *Olympus Mons,* rising 26 km from a base 650 km wide and with a caldera about 80 km wide. Olympus Mons is the largest known volcano in the entire Solar System, its volume being nearly one hundred times greater than that of Mauna Loa, the largest volcano on Earth.

It is very likely that the volcanic plains of the northern hemisphere are constituted by floods of basalt. Their features are indistinguishable from terrestrial counterparts, for example on Hawaii. Possibly, they are in part komatiitic (Baird & Clark, 1984). Crater counting indicates that the extrusion took place over much of geologic time (Neukum & Hiller, 1981). Also in view of the basaltic composition of the dust samples analysed by the Viking landers, which may be taken to represent composite samples that monitor the exposed martian crust, the whole crust of Mars is thought to be basaltic (Carr, 1981).

The differentiation of a basaltic crust makes rather certain that Mars also is a layered planet, but the intermediate density of 3.934, versus 5.514 for the Earth and 3.344 for the Moon (the uncompressed densities are 3.7, 4.0 and 3.3, respectively, Taylor, 1988), makes a large metallic core impossible. Mars has no magnetic field. The martian crust has not been subjected to extensive horizontal movements, as is indicated by the undeformed craters across the surface. A single lithospheric 'plate' appears to enclose the entire planet and has likely done so since the formation of the oldest preserved surface geological units (Solomon, 1978). Significant tectonic activity has only occurred in the way of large domal upwarps, normal faults and grabens, particularly in the northern hemisphere. These tectonic features appear to be genetically related. The most conspicuous upwarp is the *Tharsis bulge,* an area as large as North America. The crust surrounding the bulge is intensely fractured. A system of huge fractures, which have been widened and shaped by erosion, gave rise to the most spectacular tectonic feature of Mars, the *Valles Marineris,* a canyon system extending eastward from the Tharsis bulge. The system is up to 7 km deep and about 5000 km long, stretching a quarter of the way around the planet. Actually, this feature had already been observed by telescopes from Earth and turned out to be the only one of the notorious *canali* claimed a century ago by Giovanni Schiaparelli and Percival Lowell, that is real – the others were optical illusions. It is thought that the canyon system was formed because of the spreading and cracking of the crust by the doming of the Tharsis bulge. Studies of the chronology of the various tectonic features of the Tharsis bulge, again based on crater densities, indicate that the uplift began some 3 billion years ago, and that the surface fracturing had tapered off by about 2 billion years ago.

It is thought that the Tharsis uplift was produced by a single mantle plume that caused a buckling and swelling of the crust. Mantle hot spots are also held responsible for the martian volcanoes, as is the case for the volcanoes on Venus and some volcanoes on Earth, for example Hawaii. Because of the absence of plate-tectonic motion, the crust remained stationary over a hot spot. This explains why the martian volcanoes are so much larger than those on Earth – although, of course, the lower surface gravity also contributed to the higher elevation. The question whether the martian volcanoes are still active, remains unanswered. No evidence of present-day eruptions exist. Crater counting points to cratering model ages as young as some 100 million years for the youngest volcanoes, such as Olympus Mons (Hartmann, 1978; Neukum & Hiller, 1981). Olympus Mons may thus very well be dormant, possibly intermittently active with intervals of many million years between major eruptions.

Messengers from Mars

Planetary geologists all over the world eagerly await the samples from the surface of Mars that will be brought to Earth by the *Mars Sample Return Mission* around the turn of the century (NASA, 1986). Probably, however, we do possess already a few martian samples. During the last years, evidence has been accumulating that a strange group of eight stony meteorites, collected from places as far apart as India and the Antarctic ice cap, represent rocks that have been blasted off Mars by giant meteorite impacts and hurled towards Earth millions of years ago. These alleged samples of Mars are collectively known as *SNC meteorites,* from the names of the towns where the first members of this group were found: Shergotta in India, Nakhla in Egypt and Chassigny in France. The SNCs (McSween, 1985, 1987) are igneous rocks of basic to ultrabasic composition and were derived by differentiation processes from a source region in the mantle of their parent body. They show a striking resemblance to basaltic lava, indicating volcanism on the world they came from. Isotopic dating indicates that they solidified about 1.3 billion years ago, contrary to all other meteorites which typically have ages around 4.5 billion years – the age of the Solar System. This means that there must have been relatively young volcanic activity on the world where the SNCs came from, which in turn implies that their world must have been large enough to maintain volcanic activity for a long period after it formed. Asteroids, thought to be the parent bodies for meteorites, are too small. The Moon has had no volcanic activity younger than 3 billion years. A larger object is required, and Mars, with relatively young volcanism, seems to be the only candidate.

An important discovery in this respect is that gases trapped in SNCs are totally different from gases found in other meteorites or in the Earth's atmosphere, but have relative abundances and isotopic compositions of Ar, Kr, Xe and N that resemble in striking detail the compositional pattern measured by the Viking spacecraft in the martian atmosphere (Bogard & Johnson, 1983). The martian origin of the SNCs received also support from the recent finding in Antarctica of meteorites composed of anorthositic breccia, which are unquestionably pieces of the Moon blasted off by impacts and hurled to Earth (Ryder, 1987). These meteorites of lunar origin have weakened the objections that a meteoritic impact cannot blast a piece of rock away from the gravitational field of Mars. If the SNCs are indeed of martian origin – as is nowadays generally assumed, but can, of course, only be verified by the return of samples of known martian material – then they presumably represent samples of some of the large lava flows that constitute the volcanic plains, probably in the Tharsis area.

Table 3. Radiogenic and fissiogenic isotopes produced by extinct short-lived radionuclides, observed in excess in planetary bodies.[1]

Radiogenic/ fissiogenic isotope	Extinct parent	Half-life in 10^6 year	Decay or fission
$^{142}Nd^{2}$	^{146}Sm	103	alpha
$^{131\text{-}136}Xe^{2,3}$	^{244}Pu	82	alpha & fission
$^{129}Xe^{2,3,4}$	^{129}I	15.7	beta
$^{107}Ag^{2}$	^{107}Pd	6.5	beta
$^{53}Cr^{2}$	^{53}Mn	3.7	beta
$^{26}Mg^{2}$	^{26}Al	0.75	beta

[1] After Podosek & Swindle (1988).
[2] Excess observed in meteorites.
[3] Excess observed in surface-correlated gases in lunar samples, and in terrestrial atmosphere.
[4] Excess observed in martian atmosphere, and in terrestrial mid-ocean ridge basalts and deep gas wells.

Crustal evolution

All inner planets and the Moon have thus crusts which differ substantially from the planetary bulk composition. This appears to apply also to most or all solid satellites of the Jovian planets. Although there is much diversity between the planetary objects, some general principles of crustal evolution are apparent (Taylor, 1989). First, on all solid planets and moons planetary differentiation consequent upon planetary-wide melting gave rise to a magma ocean (e.g., Matsui & Abe, 1986), from which a *protocrust* differentiated within a relatively short time span after the accretion – at most some millions of years. An example is the anorthositic crust in the lunar highlands. On Earth nothing has been preserved of the protocrust due to later geological activities, but for various reasons (Taylor, 1989) it is unlikely that this early crust was anorthositic. Probably, the terrestrial protocrust was basaltic or komatiitic (Taylor & McLennan, 1985). Later in planetary history a *secondary crust* developed as a result of partial melting in the planetary interiors. Secondary crusts are composed of basalt or komatiite, the primary melts from peridotitic mantles. Examples of secondary crusts include the lunar maria basalts, the basaltic crust of Venus and Mars, and the oceanic crust on Earth. The generation of secondary basaltic crust on Earth and probably also on Venus and Mars is still going on. Eucritic meteorites, which have a basaltic composition, are probably another example of the devel-

Table 4. Hypothetical atmospheres of the terrestrial planets in absence of weathering processes and life, and without escape to space.[1]

	Venus	Earth	Mars
CO_2	96.5	98	98%
N_2	3.4	1.9	1.7%
O_2	trace	trace	trace
Ar	0.004	0.019	0.085%
H_2O (liquid)[2]	~1000	3000	~100 meter
total pressure	~90	~70	~2

[1] Modified after Pollack & Yung (1980); Squyres (1984); Prinn & Fegley (1987); Morrison & Owen (1988).
[2] Averaged over the whole surface of the planet.

opment of a secondary crust on a differentiated asteroid. On Earth, the continental crust can be defined as a *tertiary crust,* developed through processes of melting and differentiation from the secondary basaltic crust after its chemical modification by reactions with ocean water. The terrestrial continental crust is probably the sole example of a tertiary crust in the Solar System.

Planetary perspective

The inner planets Earth, Venus, Mars and Mercury, and also the Moon, have much in common when compared to the giant outer planets. Their sizes and distances to the Sun are of the same order of magnitude. All are primarily composed of cosmically rare silicates and metals, with a roughly chondritic bulk composition. Except for the Moon, they formed simultaneously by accretion from the primordial solar nebula. As for the origin of the Moon, the currently most widely accepted hypothesis is that the satellite formed during a late stage of the Earth's accretion by coalescence of debris emplaced in earth-orbit as the result of either a single impact on the primitive Earth of a body the size of Mars, the 'giant impact hypothesis' (Benz et al., 1987; Ryder, 1987; Stevenson, 1987), or a series of collisions with much smaller high-velocity bodies (Ringwood, 1989). According to the first model most of the earth-orbiting debris was derived from the impactor, whereas the second model requires that most of it was derived from the Earth's mantle. Isotopic age studies indicate that the formation of the whole Solar System was completed about 4.5 billion years ago. All inner planets underwent melting and differentiation in the very beginning of their history. Isotopic anomalies of a number of elements (Table 3) in meteorites, lunar rocks, martian and terrestrial atmospheres, and terrestrial well gases and mid-ocean ridge basalts indicate that extinct radionuclides of short half-life were still present when the planetary bodies formed. This implies that the time-interval between the end of nucleosynthesis in our part of the galaxy and the formation of the planetary bodies, including the differentiation into metallic and silicate phases and the subsequent consolidation, must have been relatively short; the very short half-life of aluminium-26, only 750,000 years, indicates that at most a few million years have been involved (see, for example, the review by Podosek & Swindle, 1988). All planetary bodies suffered in the beginning of their history a heavy bombardment of meteoroids of all sizes. Heavy impact cratering seems to have been the dominant geological process in that interval, which came to an end around 3.85 billions years ago – about the time that the oldest preserved rocks on Earth were formed.

There are also dramatic differences, however. On Earth, most of the internal heat production by the decay of potassium, thorium and uranium, and possibly also remnants of the primordial accretional and core-separation heat, are released through plate-tectonic processes. In the smaller mass of Mars, on the other hand, the internal heat finds a way out chiefly through mantle plumes and hot-spot volcanism. On Venus, plate-tectonic processes may have been operative, but appear to have come to a standstill or to have become weak billions of years ago, giving way to mantle plumes and hot-spot volcanism as the only or dominant internal geological process. The Moon and probably also Mercury are dead planets for the last 3 billion years. The internal heat of these bodies is exclusively released through conduction.

Of the three inner planets with an atmosphere, only Earth has abundant liquid H_2O and a biosphere. These two facts are obviously related, since life is totally dependent on water. Liquid H_2O requires that the surface temperature be kept within rather narrow limits. Venus is too hot, Mars is too cold, and only Earth is just right. Planetary temperature is primarily determined by the distance to the Sun. If the Earth were closer to the Sun, it would probably develop a runaway greenhouse and become a hothouse like Venus. But if the Earth were to be moved farther from the Sun than its present orbit, this would not necessarily produce the refrigerator we know Mars is. Our planet could stay warm by introducing larger quantities of CO_2 in the atmosphere. Thus the problem with Mars is not just its greater distance to the Sun – Mars got into trouble for some other reason.

Strong evidence exists, although circumstantial, that Mars once had a more massive atmosphere than the present one (Table 4). Firstly, there is the observational evidence for transient liquid H_2O early in martian history, which is impossible under the present conditions of low temperature and a very thin atmosphere. Secondly, the relative abundances of the primordial (excluding radiogenic isotopes) noble gases Ne, Ar, Kr and Xe in the martian atmosphere are the same as those on Earth. This suggests that the initial relative abundances of noble gases, CO_2 and N_2 were also similar; as we have an estimate of the terrestrial ratio of total $CO_2 + N_2$ to Ne on Earth (the terrestrial value of total CO_2 includes the estimated amount tied up in rocks, the biosphere and ocean water), while the amount of atmospheric Ne on Mars is also known, it is possible to calculate how much CO_2 and N_2 should be present on Mars in order to obtain the same ratio as on Earth. Thirdly, from the enrichment of $^{15}N/^{14}N$ over the terrestrial value it is possible, knowing the escape velocity, to calculate the original N_2 abundance. Both calculations lead to the conclusion that Mars once had an atmosphere with a surface pressure equal to about twice the sea level pressure of the present terrestrial atmosphere (Carr, 1986). The abundances of the Ar isotopes and the old age of the river valley systems (about 4 billion years) indicate that this dense atmosphere was lost early in martian history. Possibly, the depletion was the consequence of 'atmospheric erosion' due to the heavy impacting of meteoroids that ravaged all inner planets until or around 3.9 billion years ago (Cameron, 1983; Prather, 1984; Watkins & Lewis, 1986). One might also speculate that the depletion was caused by the giant impact which some authors (Wilhelm & Squyres, 1984) hold responsible for the development of the hemispheric dichotomy of the martian surface. Anyhow, the rate of removal of the planetary atmosphere by the heavy impacting should have been much more efficient on Mars than on Venus and Earth, because of its smaller size and lesser gravity (the escape velocities for Venus, Earth and Mars are 10.3, 11.2 and 5.0 km/sec, respectively, Cameron, 1983).

Because of the greenhouse effect of the denser CO_2 atmosphere, the surface temperature was above the freezing point of water over much or all of the Martian surface, so that hospitable environmental conditions prevailed. In that time, until about 4 billion years ago, rivers were flowing over the martian surface, which left their mark in the branching drainage systems of runoff channels still preserved in the landscape. We may even consider the possibility that life obtained a foothold on Mars as it did on Earth in the same time, but that the biota were wiped out early in the planetary history as the environmental conditions changed to their present state. One of the priorities for future Mars missions will obviously be to search for signs of ancient life in the regions composed of the oldest sedimentary successions (NASA, 1986).

If life indeed developed in the beginning of Mars' history, only to die out as the environmental conditions broke down, then the disappearance of martian biota would, according to the above scenario, ultimately be due to the impacts of meteoroids. This catastrophe would thus somewhat resemble the biotic crises and mass-extinctions in the paleontological record on Earth that punctuate the history of our planet. Some of these crises (maybe all major ones ?) are nowadays likewise attributed to giant impacts. Contrary to Mars, however, the terrestrial biosphere has so far managed each time to recover. Each mass-extinction triggered an evolutionary explosion, with a spectacular adaptive radiation of the surviving species and a rapid development of new life forms (Stanley, 1987). Impacting meteoroids may also have played an essential role in the origin of life. Meteorites contain a variety of complex organic molecules, formed through abiological processes in the deep cold of interstellar space (Hoyle & Wickramasingh, 1976; Goldanskii, 1986). During the high meteoritic flux early in the history of our Solar System, such abiogenic organic compounds could have been deposited in great abundance on the planetary surfaces. It is speculated that these chemicals provide the building blocks that underlay, under appropriate environmental conditions, on any planet the chain of reactions leading to the assembly of life. If so, the interplanetary chunks of rock are messengers of both Life and

Death, playing in the planetary systems the role of Shiva in Hindu mythology – destructor, but also creator and renewer of Life.

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