CHAPTER 2 MARS

Mars is called the “Red Planet” due to the occurrence of oxidised iron on its surface. It is a terrestrial planet and thus shares a common structure and surface morphology as the Earth. Particularly its environmental conditions are the most earth-like of all terrestrial planets [*Spohn et al.*, 1998]. At present, it is a desert-like planet characterised by strong winds and regularly occurring dust storms. Its atmosphere is described as thin and rich in carbon dioxide (Table 5). Mars is half the size of the Earth and its geological evolution has been different (Table 4).

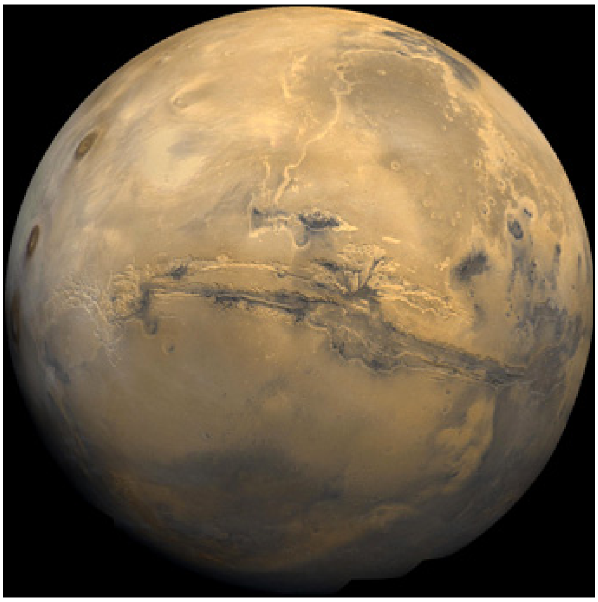

Figure 1: Mars with the huge canyon system Valles Marineris in the centre and the Tharsis volcanoes at the western edge (Viking Orbiter Image/NASA).

Table 1: Parameter of Mars and Earth in comparison.

Parameter	*Mars*	*Earth*
Equatorial radius [km]	3396	6378.1
Mass [10^{24} kg]	0.642	5.973
Mean density [kg/ m³]	3.934	5.517
Surface gravity [m/ s^2]	3.711	9.807
Moons	2 (Phobos, Deimos)	1

However, there is evidence for water that once flowed over its surface [*Carr*, 1979] indicating a different warmer, wetter Mars with a denser atmosphere than what is observed today. Several ESA and NASA-missions (orbiters and landers, Table 2) provide excellent data of its surface enabling a detailed geological analysis and comparisons with Earth.

Table 2: Missions relevant to Mars Surface Science from 1971-current. Modified after *Bell* (2008).

Launch/Status	*Mission*	*Nation/Agency*	*Failure*
1971-72	Mariner 9	NASA	
1973	Mars 4	UdSSR	failed orbit
1973	Mars5	UdSSR	
1976-80	Viking 1 and 2	NASA	
1988	Phobos-1, Phobos-2	USSR	Phobos-1 failed; Phobos-2 reached Mars, survived for several months
1992	Mars Observer	NASA	failed: lost prior to Mars arrival
1996-2006	Mars Global Surveyor (MGS)	NASA	
1996	Mars 96	USSR	launch vehicle failed
1996-1997	Mars Pathfinder (MPF)	NASA	
1998	Nozomi	Japan	failed: no orbit insertion
1998	Mars Climate Orbiter	NASA	lost on arrival
1999	Mars Polar Lander	NASA	lost on arrival
1999	Deep Space 2 Probes	NASA	carried by MPL, failed: no signal after release
2001-current	Mars Odyssey	NASA	
2003-current	Mars Express (MEX) Orbiter	ESA	
2003	Beagle 2 Lander	ESA	carried by MEX, failed landing
2003-current	Mars Exploration Rovers (MER)	NASA	
2005-current	Mars Reconnaissance Orbiter (MRO)	NASA	
2007	Phoenix Lander	NASA	

Table 3: Modelled age of the stratigraphic Martian epochs and periods modified after *Jaumann* (2003). Stratigraphy after *Tanaka* (1986), Two age models are present, one is based on Neukum-data shown by N and the other one is based on Hartmann-data shown by H [*Hartmann and Neukum*, 2001].

Stratigraphy	*Age [Ga] before today*	*Age [Ga] before today*
Early Noachian	4.65 - 3.95	
Middle Noachian	3.95 - 3.8	
Late Noachian	3.8 - 3.7	
Early Hesperian	3.7 - 3.6	
Late Hesperian	3.6 - 3.1 (N)	3.6 - 2.9 (H)
Early Amazonian	3.1 (N) - 2.1 (N)	2.9 (H) - 1.4 (H)
Middle Amazonian	2.1 (N) - 0.55 (N)	1.4 (H) - 0.3 (H)
Late Amazonian	0.55 (N) - today	0.3 (H) - today

Table 4: Geological processes at different Martian epochs and periods modified after *Scott and Tanaka* (1986).

Stratigraphy		***Processes***
Noachian	Early	• high meteorite flux and intense bombardment of Martian surface • rugged basement rocks, mountains • Argyre impact basin Planitia formed
	Middle	• northern lowlands developed • decreasing impact rate • emplacement of volcanic and impact-breccia plateau material • extensive faulting formed Claritas, Coracis, Acheron, Melas, and Nectaris Fossae
	Late	• widespread resurfacing and intercrater filling partly subdued older surfaces • formation of large ridges in Terra Sirenum and Noachis Terra • faulting radial to Syria Planum formed Ceraunius, Tempe, Mareotis, and Noctis Fossae • highland surfaces channelled and etched • extensive fluvial erosion by valley networks
Hesperian	Early	• eruption of lava flows (later ridged) in intercrater and lowland plains and in Lunae Planum • long, tongue-shaped lava flows emplaced east of Alba Patera and at Tempe Terra • faulting and rifting dominantly radial to Syria Planum-Pavonis Mons formed Valles Marineris and Ulysses, Memnonia, Sirenum, Icaria, Thaumasia, and Fortuna Fossae • widespread degradation and burial of most older cratered terrain by lava flows in lowland region • fluvial erosion by valley networks
	Late	• extensive lava flows erupted from sources at Tharsis Montes, Alba Patera, Ceraunius Fossae, Tempe Terra, and Syria Planum • deposition of layered material in chasmata • emplacement of lavas or sediments in northern Acidalia and Arcadia Planitiae • faulting of Noctis Labyrinthus • formation of chaotic terrain and large outflow channels north and east of Valles Marineris, extending into flood plains in Chryse Planitia and in southern Acidalia Planitia

Amazonian	Early	• extrusion of younger lava flows on and surrounding the Tharsis Montes and Alba Patera and in lowlands of Amazonis, Arcadia, and Acidalia Planitiae • formation of oldest Olympus Mons aureoles
	Middle	• flows emplaced around Tharsis Montes • eruption of lavas continued in Amazonis Planitia • deposition of soft-appearing material in the Memnonia Sulci and Gordii Dorsum areas
	Late	• ribbed debris aprons formed on northwest flanks of the Tharsis Montes and Olympus Mons, accompanied by minor volcanism • channel deposits in western Amazonis Planitia • deposition of plains material in Arcadia Planitia • deposition of soft material continued in Medusae Fossae and Amazonis Sulci areas

2.1 MORPHOLOGICAL STRUCTURE AND COMPOSITION

The crust of Mars is divided between the northern lowlands (Fig. 2), which are mainly extended plains covered by lava flows and sediments (Hesperian, Amazonian) and the ancient, rough, heavily cratered southern highlands (~60 % of the surface; Fig. 2), which in turn represent the oldest formations (Noachian, Hesperian ages; Table 3; *Tanaka et al.*, 1992). The northern lowlands appear smooth, hardly cratered, and partially covered by lava; both also show indications of aeolian, fluvial, and glacial processes affecting the surface. The northern part is thought to consist of an andesite-like material, while the composition of the southern part is basaltic [*Bandfield et al.*, 2000]. The South Pole (Noachian, Hesperian) shows indications of glacial and fluvial processes, whereas the North Pole is characterised by aeolian, glacial, and fluvial morphologies (Amazonian, *Head et al.*, 2001), which form dune fields and surface mantling. Both polar regions exhibit a huge amount of layered deposits consisting of interbedded strata of ice and dust which, appearing light and dark, are covered by H_2O and CO_2 ice and frost at the poles themselves.

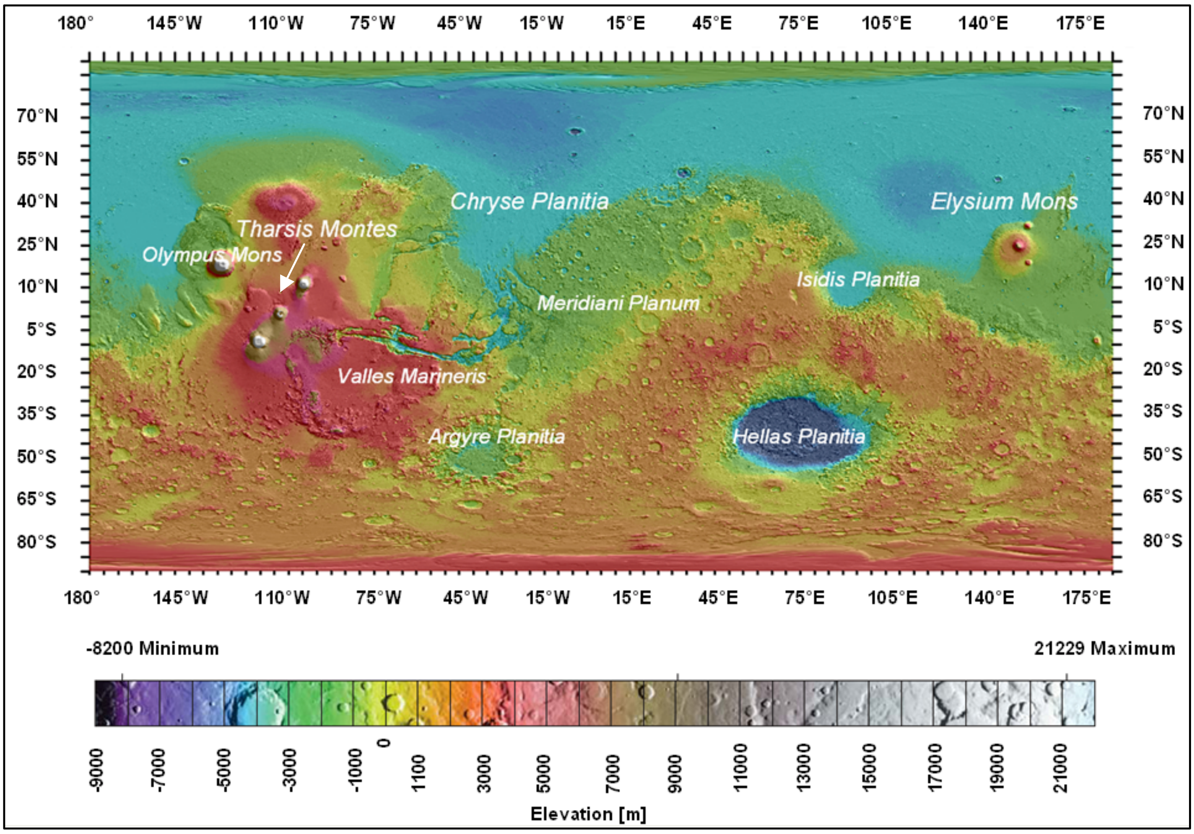

Figure 2: *MOLA* topographical map of Mars. Note the dichotomy boundary visible by elevation that separates the northern lowlands (blue, green colours) from the southern highlands (yellow to red colours); modified after topographical map M 25M RKN by U.S. Geological Survey 2003.

Mineralogical observations by OMEGA [*Bibring et al.*, 2006] suggest that there were three sequential eras (Fig. 3): (i) a non-acidic aqueous alteration, traceable by phyllosilicates (*"PHYLLOSIAN"* era, (ii) an acidic aqueous alteration, traceable by sulphates (*"THEIIKIAN"* era; and an atmospheric aqueous-free alteration, traceable by ferric oxides (*"SIDERIKIAN"* era). The transition between the first two eras (Fig. 3) is supposed to have taken place in the early age of Martian, when the climate changed from an alkaline, humid to an acidic environment. The outgassing of volatiles, including sulphur, and the overall volcanic activity might have driven this climate change. *Bibring et al.*, 2006 suppose that the environment might have always been tenuous, cold and dry, except for transient periods, if phyllosilicates were formed below the surface. Conversely, if phyllosilicates formed at or close to the surface they suppose the early Martian atmosphere to have been dense. The transition to an acidic environment would then have occurred with a rapid decline in atmospheric pressure (Table 6). Rapid lowering of the atmospheric pressure was assumed either due to the rapid drop of the internal dynamo and its resulting magnetic shield that favoured atmospheric sputtering or to the heavy bombardment and the Martian present low mass. Several processes are being considered here: (i) the heavy bombardment; (ii) Mars' low mass, which lead to the escape of the atmosphere [*Bibring et al.*, 2006]; and (iii) the rapid drop of the internal dynamo and its resulting magnetic shield. They assume that the presence of phyllosilicates indicates an era of habitable conditions in which biochemical developments might have taken place.

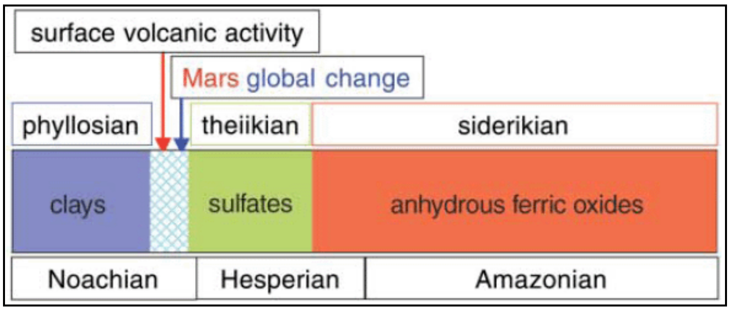

Figure 3: Sketch demonstrating the Martian alteration history suggested by *Bibring et al.* (2006). Phyllosilicates form first -in the Noachian. After the volcanic activity set in on the Martian surface and the climate underwent a global change triggered by volcanic activity, drop of the magnetic shield and decline of the atmospheric pressure. Both the availability of water and its pH decreased, resulting in the precipitation of sulphates and afterwards of anhydrous ferric oxides.

2.2 ATMOSPHERE AND CLIMATE

On Mars today, the soils and rocks become much warmer by absorption than the air, due to the very thin atmosphere. The atmosphere predominantly consists of CO_2 (95.3 %, Table 5). Nitrogen (2.7 %) and argon (1.6 %) are only minor amounts. The mean Martian surface temperature is -53°C but at the equator, temperatures may reach 27°C (Table 6) whereas the lowest temperatures are at -140°C. On Mars, the current mean atmospheric pressure is 6 mbar which is less than 1 % at sea level on Earth (Table 6). These low temperatures and pressures enable the deposition of carbon dioxide from the atmosphere at temperatures of -123-133°C, which is accomplished at the poles *Reiss* (2005). However, liquid carbon dioxide on Mars can be excluded.

Table 5: Composition of the atmosphere.

Composition of the Atmosphere	***[%]***
Carbon dioxide (CO_2)	95.32
Nitrogen (N_2)	2.7
Argon (Ar)	1.6
Oxygen (O_2)	0.13
Carbon monoxide (CO)	0.07
Water (H_2O)	0.03

Table 6: Atmospheric parameters of Mars and Earth.

Parameter	***Mars***	***Earth***
Mean atmospheric pressure [mbar]	6.5	1013
Minimum surface temperature [°C]	-140	-80
Mean surface temperature [°C]	-63	13
Maximum surface temperature [°C]	27	45

The Martian day is 25 hours 37 minutes and its year consists of 669 days (Table 7) resulting from its more eccentric orbit around the sun as the Earth. The aerocentric length of the sun (L_S called solar longitude) is a measure for the constellation of Mars in its orbit and thus defines its seasons. There, the solar longitude (Table 7) is the angle between

Mars, sun and the autumnal equinox.

Table 7: Duration of seasons on Mars.

L_S	*Northern Hemisphere*	*Southern Hemisphere*	*Number of Days on Mars*	*Number of Days on Earth*
0° - 90°	Spring	Fall	194	199
90° - 180°	Summer	Winter	187	183
180° - 270°	Fall	Spring	143	147
270° - 360°	Winter	Summer	154	158
			669	**687**

Today, there is no liquid water on the Martian surface because the atmosphere is thin and temperatures are low (Sect. 2.1, Table 6) but for instance relict channels and valley networks indicate hydrological conditions were different from today's [*Baker et al.*, 1992, *Jaumann*, 2003]. The presence, age and distribution of the valley networks (3.8 Ga) on Mars clearly shows that the process of erosion changed compared to mid-to-late Martian history, which indicates a long drawn-out climate change [*Carr and Clow*, 1981]. Valley networks are restricted to ancient, heavily cratered terrain and may have been formed by liquid water at a time when Mars was warmer and wetter and had a dense atmosphere.
After the heavy bombardment, conditions might have permitted running water to create valleys [*Malin and Edgett*, 1999]. Outflow channels are supposed to have been eroded fluvially as a consequence of the catastrophic release of water [*Baker et al.*, 1992]. Most of them emanate from chaotic terrain [*Sharp*, 1973a], which is heavily collapsed terrain and assumed to be their source region. Outflow channels reveal another Martian climate of hydrological conditions in the Late Hesperian or Amazonian when they were formed [*Tanaka*, 1986]. In addition, sapping rather than rainfall was assumed for Mars [*Fanale et al.*, 1992]. Further information can be found in *Reiss* (2005) and more details are given in *Kieffer et al.* (1992); *Carr* (1996b) and *Carr and Head* (2003).

2.2.1 Weathering

The term weathering describes physical and chemical processes in rocks and minerals exposed to a hydrosphere and atmosphere. Thus, subaeric (in contact with the atmosphere) and subaquatic (underwater) can be distinguished [*Matthes*, 2001]. Locally, weathering products form soils and, after transport, sediments, and sedimentary rocks. The basic material is derived from magmatic, metamorphic, or old sedimentary rocks whose material has already undergone a sediment forming process.
There is mechanical (physical) and chemical weathering. Purely mechanical weathering (caused by heat, frost, salt) will disintegrate exposed rocks into loose bulk material without any change in chemical composition. Mechanical weathering predominates in cold climates, whereas chemical weathering prevails in climates where moisture is present. From Earth observations, it is known that quartz grains of sand and silt size, mica (muscovite), clay minerals (montmorillonite) and various soluble carbonates as well as iron oxides are the end products of weathering [*Greeley and Iversen*, 1985]. On Earth,

both processes mentioned above are generally involved in weathering, whereas on Mars mechanical weathering dominates due to the scarcity of moisture. On Mars, unweathered olivine, pyroxene, and feldspar were found throughout the equatorial region indicating that no extensive aqueous weathering has occurred there [*Christensen et al.*, 2000b; *Christensen et al.*, 2003; *Hoefen et al.*, 2003; *Christensen and Ruff*, 2004].

2.2.2 Water

There are adequate indications that liquid water existed early on the Martian surface, including outflow channels and valley networks (Table 4, *Carr*, 1979). Due to the size of the outflow channels, large amounts of water might have flown on the surface (Sect. 1, 2.2). The morphology and size of the valley networks indicate that liquid water must once have been stable on the surface [*Wallace and Sagan*, 1979; *Carr*, 1983]. Some valley networks show evidence of groundwater sapping and melting of ice [e.g. *Carr and Clow*, 1981]. On the other hand, clearly defined dendritic systems were probably caused by precipitation and runoff [*Masursky et al.*, 1977]. For estimations of the entire amount of water released at the surface, see *Carr* (1986); *McKay and Stoker*, (1989).
Under the current Martian conditions, water is theoretically is stable in three phases of which the solid and the gaseous state offer long-lasting stability (Fig. 4). The triple point of water (Fig. 4) indicates that a temperature of > 273 K and a pressure of > 6.1 mbar are the minimum physical conditions.

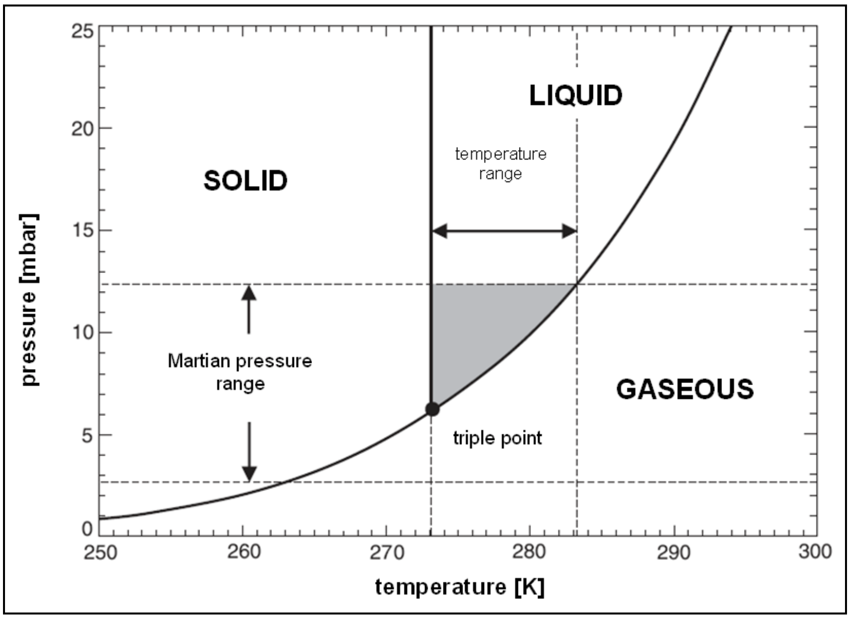

Figure 4: Phase diagram of water modified after *Haberle et al.* (2001); *Reiss* (2005). The cross-hatched area represents the range of pressures and temperatures where liquid water could form on Mars.

2.3 Geology

2.3.1 Aeolian processes

The most active and probably only surface-modifying process on Mars is the wind action. For 4.7 Ga (Table 3) aeolian processes have played a significant role in its surface evolution. Most surface morphologies indicate wind affection (e.g. dune fields, albedo patterns, yardangs, drifts of fine-grained material). Large parts of the cratered uplands and smooth terrains in the polar regions are presumably composed of windblown particles [*Greeley et al.*, 1992].

Aeolian processes only occur if there are adequate supplies of loose particles and winds of sufficient strengths to move them. Dust storms on Mars are rare but they greatly influence the appearance of the surface. However, extensive sedimentary deposits are rather due to long-term processes like atmospheric dust transport rather than dust storm events [*Greeley et al.*, 1992]. The most common aeolian morphologies are wind streaks, which are formed by wind erosion as well as deposition. Sand dunes [e.g. *Tirsch et al.*, 2008] and drifts have been observed in all regions, indicating aeolian deposits (Fig. 5). There are also features like yardangs and possible ventifacts. Windblown sand and dust sometimes mantle wide areas that are not of aeolian origin.

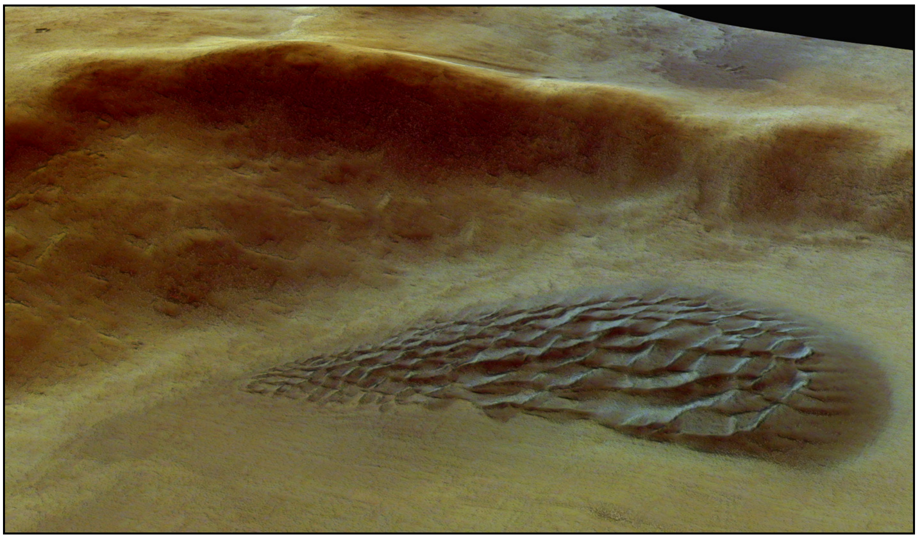

Figure 5: Dark dune field in the Argyre Planitia impact crater (*HRSC* perspective view). The extent of the dune field is 7 by 12 km.

The atmospheric density of Mars, which is low compared to Earth (Table 1) complicates aeolian transport because much higher wind velocities are required to move particles [*Greeley et al.*, 2001]. On Mars, aeolian transport mainly affects sand-sized particles of a grain size of 100 µm via saltation [*Greeley et al.*, 1980]. The minimum wind velocity to move these particles depends mostly on atmospheric density [*Greeley et al.*, 1992]. There the threshold velocity at the Martian low pressures (Table 1) is ten times higher than on Earth. Thus under the current atmospheric conditions a velocity of 25-30 m/s is required to move the abovementioned grain sizes but mostly they are <10 m/s [*Greeley et al.*, 1980]. Most grains that are transported are the size of coarse sand, whereas the transport of larger of finer grain sizes (e.g. dust) requires higher wind velocities. In fine particles such as dust, this is can be explained by interparticle forces (cohesion) and aerodynamic effects, whereas in larger particles it is explained by their higher mass [*Greeley et al.*, 2001]. However, during certain dust storm events the required velocities to move particles are attained.

Wind erosion on Mars operates by either deflation or abrasion [*Greeley et al.*, 2001]. Deflation implies movement of relatively loose materials such as unconsolidated aeolian sediments or grains of rock weathering products. It produces wind-stripped surfaces and depressions such as blowouts and is most efficient in regions that are heavily affected by physical weathering and show no vegetation [*Murawski*, 1991]. On the other hand, abrasion occurs when rock particles are dragged over or hurled against a surface, causing fragmentation. Thus, rocks may feature pits, flutes and grooves on a millimetre to centimetre scale [*Greeley et al.*, 2001]. On a larger scale (cm-m), rocks may be faceted and eroded into ventifacts, while yardangs are formed by a combination of abrasion and deflation on a metre to kilometre scale.

2.3.2 Volcanism

Volcanic processes were present throughout Martian history [*Carr*, 1973; *Greeley and Spudis*, 1978; *Mouginis-Mark et al.*, 1992; *Tanaka et al.*, 1992; *Greeley et al.*, 2000; Sect. 2.1]. Volcanic activity such as magma generation and extrusion indicates processes in the interior. The evolution of the interior may be traced from the age, location and type of volcanic material. Material was released either by central volcanoes (e.g. Tharsis; Fig. 6, Table 4) or on plateaus as flood lava [*Head et al.*, 2001].

Much of the ancient Martian crust is volcanic [*Christensen et al.*, 2000b] and primarily of basaltic composition (surface type 1) in the ancient southern highlands and of andesite-like composition (surface type 2) in the northern lowlands [*Bandfield et al.*, 2000; *Christensen et al.*, 2001a; *Ruff and Christensen*, 2007]. For identifying volcanic materials, the presence of feldspar is required [*Matthes*, 2001]. On Mars, these are feldspars (anorthite-rich plagioclases, Sect. 3.2.2) identified by TES (Sect. 3.1.5, 3.2.2). Spectrometric instruments like OMEGA (Sect. 3.1.7, 3.2.2) also confirm the presence of mafic minerals such as olivine and pyroxene [e.g. *Mustard et al.*, 2005; *Poulet et al.*, 2007; Sect. 3.2.2]. Although morphological features such as flows and vents lack in the ancient Martian crust to indicate a volcanic origin, comparison with Earth and Moon [*Stöffler and Ryder*, 2001] as well as modelling of Mars [*Spohn et al.*, 2001] shows that magmatic activity occurred.

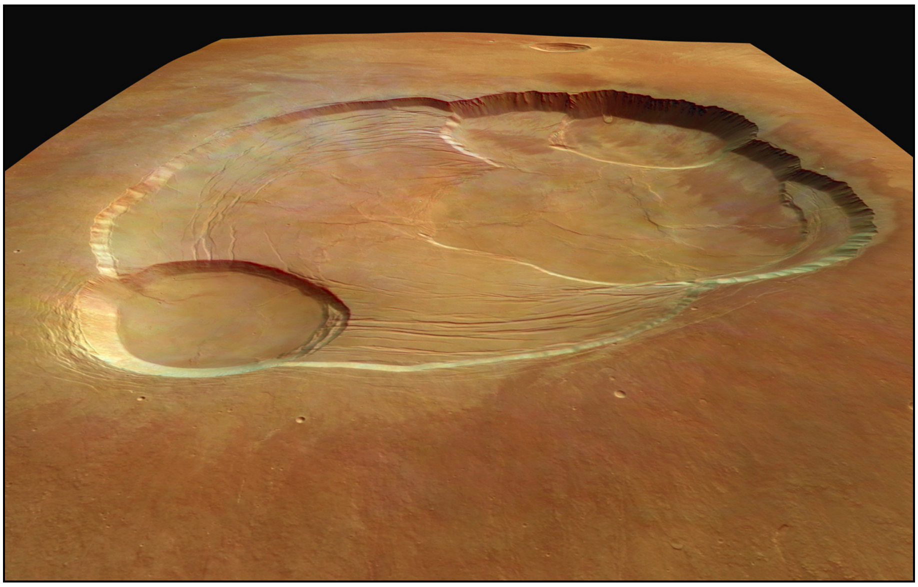

Figure 6: Olympus Mons caldera. Olympus Mons is the highest volcano in the solar system (*HRSC* perspective view). Image is 102 km across.

2.3.3 Tectonics

There is evidence of tectonic processes in the form of various structures indicating brittle extensional (grabens, rifts, tension cracks, troughs) and compressional (wrinkle ridges, lobate scarps) processes [*Head et al.*, 2001]. According to *Head et al.* (2001), their relative ages and causative processes can be analysed by structural mapping and crater counts. Besides, lithospheric stress is modelled based on topographic and gravity information [*Golombek and Banerdt*, 1999]. Stress in the lithosphere of a planet on a global scale may be caused by thermal contraction, despinning or polar wandering but it is not clear whether these processes occurred on Mars as the tectonic evidence for them is too weak [*Banerdt et al.*, 1992].
The main tectonic structures are the Valles Marineris and the adjacent Tharsis region with its volcanoes (Fig. 2). There are numerous fractures and grabens (Fig. 7) which mostly extend radially from the Tharsis bulge [*Carr*, 1974; *Blasius et al.*, 1977].

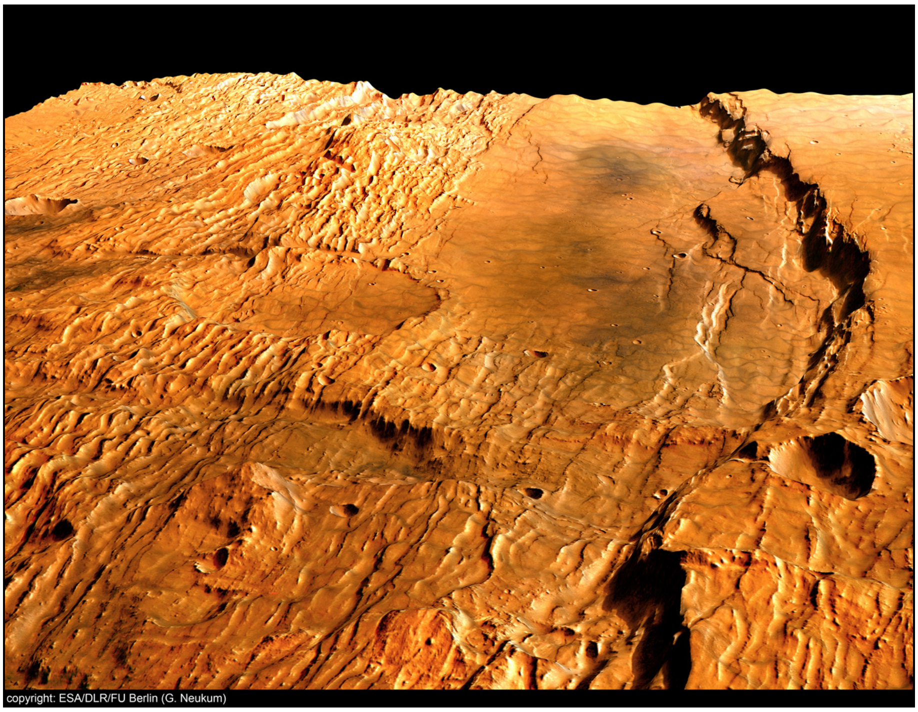

Figure 7: Claritas Fossae – fracture zones in the Thaumasia region (*HRSC* perspective view).

These features are supposed to have been formed by crustal extension while the Tharsis volcanism was active [e.g. *Carr*, 1974; Table 4]. Erosional processes (subsurface water or ice) might then have enlarged them. Their close proximity to the eastern outflow channels in the Xanthe Terra regions also argues for fluvial processes being involved in the Valles Marineris formation. Chaotic terrains are assumed to be the source regions of the outflow channels. Their formation is associated with volcanic activity in the Tharsis region (Table 4) which released of confined sub-surface water or ice so that the surface collapsed, forming chaotic terrains [*Rodriguez et al.*, 2005].

2.3.4 Impact

The surface of Mars is dominated by impact craters (Fig. 8), especially in the ancient southern highlands (Sect. 2.1), which, with their large basins like Hellas and Argyre [*Head et al.*, 2001], are assumed to be older than surfaces exhibiting small craters in the northern lowlands (Sect. 2.1, Fig. 2). When an impact forms a crater, the surface is dug up and crustal material in the form of ejecta is relocated laterally. The ejecta deposits in younger craters provide information about the substrate and the cratering process itself. *Carr* (1996a) and *Cabrol and Grin* (2001), for instance reported on craters that served as sinks for ponded surface water as a result of groundwater release during heating and impact melt emplacement that penetrated the cryosphere.

The surface chronology is based on impact cratering method developed by *Neukum and Hiller* (1981), *Neukum*, (1983) and *Hartmann and Neukum* (2001). Those pioneered in the definition of surface chronology by an impact cratering model adapted from the moon. *Hartmann and Neukum* (2001) found that the age of young lavas from the Late Amazonian is consistent with the age of Martian meteorites, while most of the Noachian dates before 3.5 Ga. *Tanaka et al.* (1987) used the tabulation of areas resurfaced by different geological processes in diverse epochs to compute the resurfacing rate over time (Fig. 9). Many endogenous processes were apparently more active in the early period of Mars (~3 Ga, Table 4), followed by a decline.

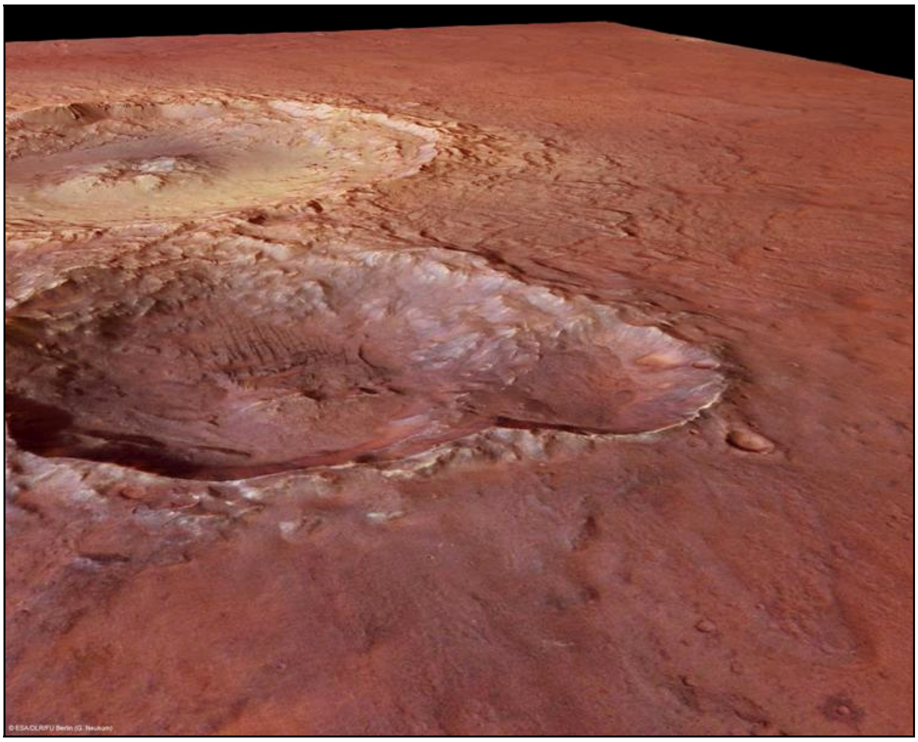

Figure 8: Tyrrhena Terra impact crater (*HRSC* perspective view).

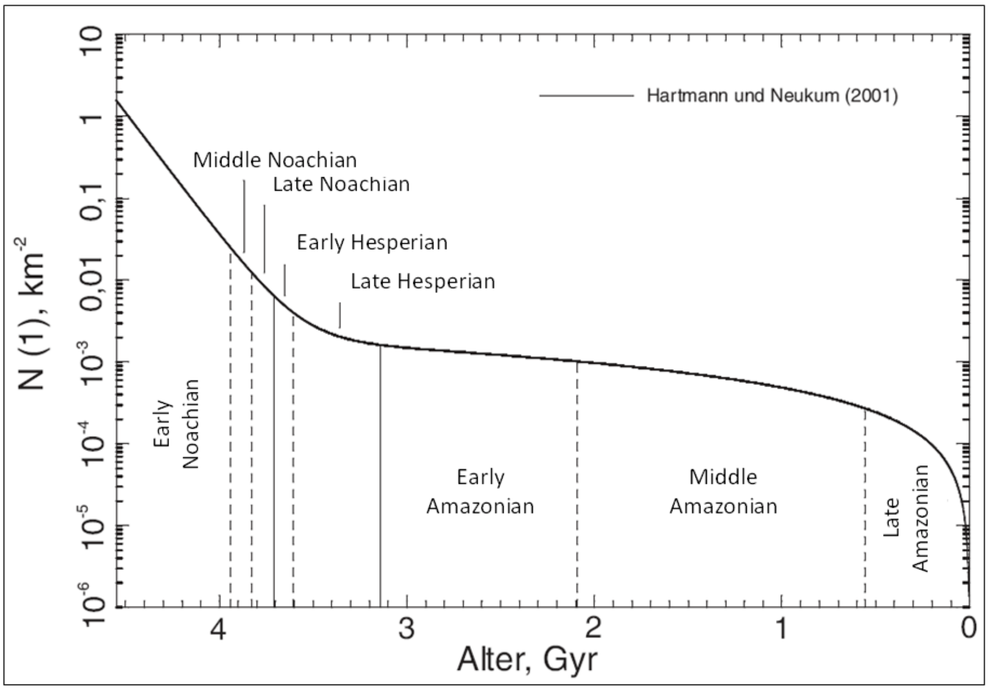

Figure 9: Martian impact cratering and chronology curve. Chronological periods and epochs after *Tanaka* (1986) and redefinition of the Lower (Early) Amazonian base crater frequency by *Hartmann and Neukum* (2001); modified after *Reiss* (2005).

2.4 Relationship between Valles Marineris, Chaotic Terrains and ILDs

2.4.1 Valles Marineris

The graben system Valles Marineris (Fig. 10, 2), named after the Mariner 9 mission, is situated in the equatorial region [*Sharp*, 1973b]. According to *Lucchitta* (1990) and *Scott and Tanaka* (1986, Table 4), its formation is dated to the Early Hesperian (Table 3, 4). From west to east, Valles Marineris extends for about 4,000 km from longitude 250°E to 320°E. The Noctis Labyrinthus trough networks are located in the west and the linear troughs and the Valles Marineris in the centre [*Blasius et al.*, 1977]. The east is characterised by wide troughs merging into chaotic terrains and outflow channels [*Lucchitta et al.*, 1992]. Some chasmata are more than 100 km in width, aggregating to form depressions that are more than 600 km wide in the central part of the graben system [*Lucchitta et al.*, 1992]. With a maximum depth of 10 km they reveal the upper Martian crust in three dimensions. Valles Marineris is one of the lowest regions of the planet.

Valles Marineris is often called the Grand Canyon of Mars, although the term canyon is misleading, as it was not carved out by fluvial erosion and uplift, as was the Grand Canyon of Arizona. As a graben [*Scott and Tanaka*, 1986], it was formed by tectonic and volcanic processes during the evolution of Tharsis (Fig. 10, Sect. 2.3.3, Table 4,). Thus, it may be properly called a graben or rift like the East-African Rift system. There are similarities between the two in the distribution of length [*Frey*, 1979]. Whereas on Earth the width and structure of continental rifts depend on the degree of crust extension, width and depth alternate on Mars [*Milanovsky and Nikishin*, 1984]. Unlike terrestrial rifts, Valles Marineris has a significantly greater width possibly because of the higher thickness of the crust or lithosphere. These differences may be due to the crust in Martian rift zones being thicker and more homogenous than on Earth.

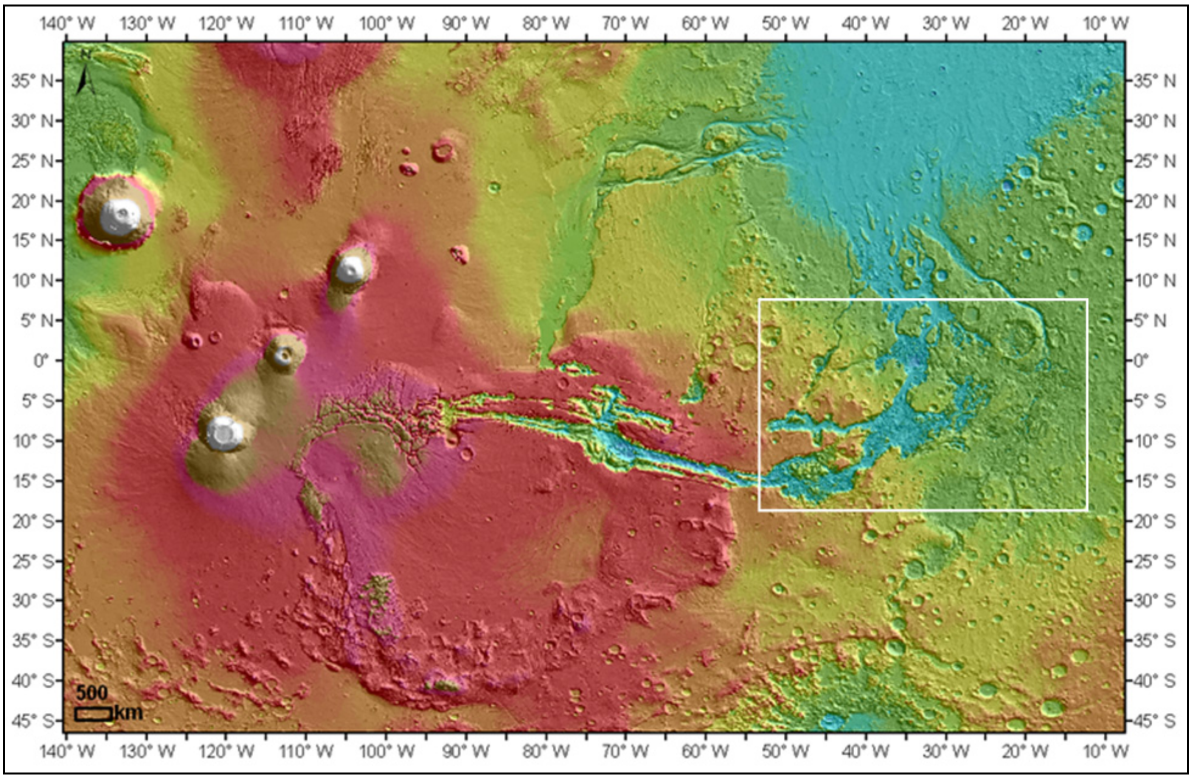

Figure 10: MOLA-map showing the extent and complexity of the Valles Marineris canyon system. For context, compare figure 2. The Valles Marineris stretches from the Tharsis region in the west to the chaotic terrains in the east. Blue marks the lowest and red the highest elevations. There are several parallel fractures; the larger ones are widened by collapse and erosion processes. White box indicates the research area (Fig. 28).

Most terrestrial rifts have many parallel en echelon faults tapering out along strike, while the rifted beds are tilted. However, the faults in Valles Marineris are widely spaced, trough ends are blunt, fault planes appear to be steep, and there are hardly any tilted beds [*Lucchitta et al.*, 1991].

2.4.2 Chaotic Terrains

Several regions in Valles Marineris exhibit chaotic terrain with ILDs exposed (Sect. 2.4.3). Chaotic terrains were dated to Late Hesperian [e.g. *Scott and Tanaka*, 1986; *Rotto and Tanaka*, 1995; Table 4, Fig. 11]. Chaotic terrains were described for the first time by *Sharp* (1973a) as structures derived from collapsed surfaces that are characterised by jumbled assemblages of large, irregular blocks occupying lowlands or depressions. Chaotic terrains concentrate in the source regions of the outflow channels east of the Tharsis bulge [e.g. *Sharp*, 1973a], i.e. within Valles Marineris chasmata and also within some craters (Fig. 2, 10, Sect. 2.4.1). Several chaotic terrains and chaos-like features (mensas and knobs) are located in areas with a high topographical gradient such as the dichotomy boundary (Sect. 2.1) –an area of slope instability [*Rossi et al.*, 2006].
Surface collapse is responsible for chaotic terrain formation [*Sharp*, 1973a]. This is caused by discrepancies between relatively incompetent strata that underlie competent strata-

inducing subsurface movements. In this instance, incompetent strata are supposed to consist of subsurface water ice in parts heated by volcanic activity or magma [*Ori and Mosangini*, 1998; *Rodriguez et al.*, 2005]. Since chaotic terrains are assumed the source regions of the outflow channels, catastrophic release of groundwater [*Carr,* 1979] likewise is included (Table 4, Sect. 2.2, 2.2.2). These catastrophic floods most likely played a role in the ILD formation and/or erosional processes that affected them.
Rossi et al. (2006) reported terrestrial analogues in the Porcupine Basin off Ireland. In addition to morphological similarity, the main analogy is the stratigraphic relationship of overlying competent and underlying incompetent fluidising strata.

2.4.3 Interior Layered Deposits (ILDs)

Interior Layered Deposits were first described by *McCauley* (1978) as light-toned layered deposits. These are exposed at several localities on Mars, which are in mainly depressions affected by chaotic terrain (Sect. 2.4.2). ILDs are located either in chasmata of Valles Marineris (Fig. 2, 10, Sect. 2.4.1) or in craters within the old cratered upland (Fig. 2). They are present in the Valles Marineris chasmata of Hebes (1.1°S/283.8°E), Candor (6.5°S/289.1°E), Ophir (4°S/287.5°E), Melas (10.3°S/287.3°E), Juventae (3.5°S/298.6°E), Ganges (7.9°S/311.9°E) and Capri/Eos Chasma (13.8°S/312.6°E). In Ius (6.9°S/274.2°E), N-Melas and Coprates Chasma (13.2°S/298.6°E) they lack [*Lucchitta et al.*, 1994]. They stand out from their surroundings by their distinct layering, morphology, eroded shape, freestanding occurrence and high albedo (Sect. 1).

As already described in Section 1, various hypotheses are being considered for the formation of ILDs. Thus, ILDs are suggested to be

- ancient deposits of glacial, lacustrine or aeolian origin that were exposed by erosion [*Malin and Edgett*, 2000];
- formed by volcanic material (subaeric pyroclastic volcanism [*Peterson*, 1981; *Chapman and Tanaka*, 2002; *Hynek et al.*, 2002]; subglacial volcanism [*Chapman and Tanaka*, 2001] or volcanic flows [*Lucchitta*, 1981; *Lucchitta et al.*, 1992; *Lane and Christensen*, 1998]), lacustrine deposits [*McCauley*, 1978; *Nedell et al.*, 1987; *McKay and Nedell*, 1988],
- spring deposits [*Rossi et al.*, 2007; 2008], or
- formed by salt diapirism [*Milliken et al.*, 2007] or a combination of several processes.

As already mentioned in Section 1, the exact age of ILDs is unknown due to the rareness of impact craters on their surfaces (Sect. 2.3.4). However, crater countings are useful not least to define an erosional age. Age dating on ILDs mostly pointed to a Late Hesperian and Amazonian age [*Lucchitta*, 1999; *Neukum et al.*, 2007; Table 3]. According to *Rossi et al.* (2008), ILDs in the eastern chaotic terrains show a young exposure age of less than 0.3 Ga (Late Amazonian, Fig. 11, Table 3). Particularly, they computed a Late Amazonian age (0.3 Ga and 0.07 Ga, Table 3) for Aram Chaos ILD (Sect. 4.1.1). For Iani Chaos ILDs (Iani 1 and Iani 3, Sect. 4.1.3), *Rossi et al.* (2008) appointed a resurfacing event around

0.1 Ga (Table 3). Overall, ILDs certainly are no older than the Late Hesperian and experienced multiple resurfacing up to the Late Amazonian (Fig. 11).

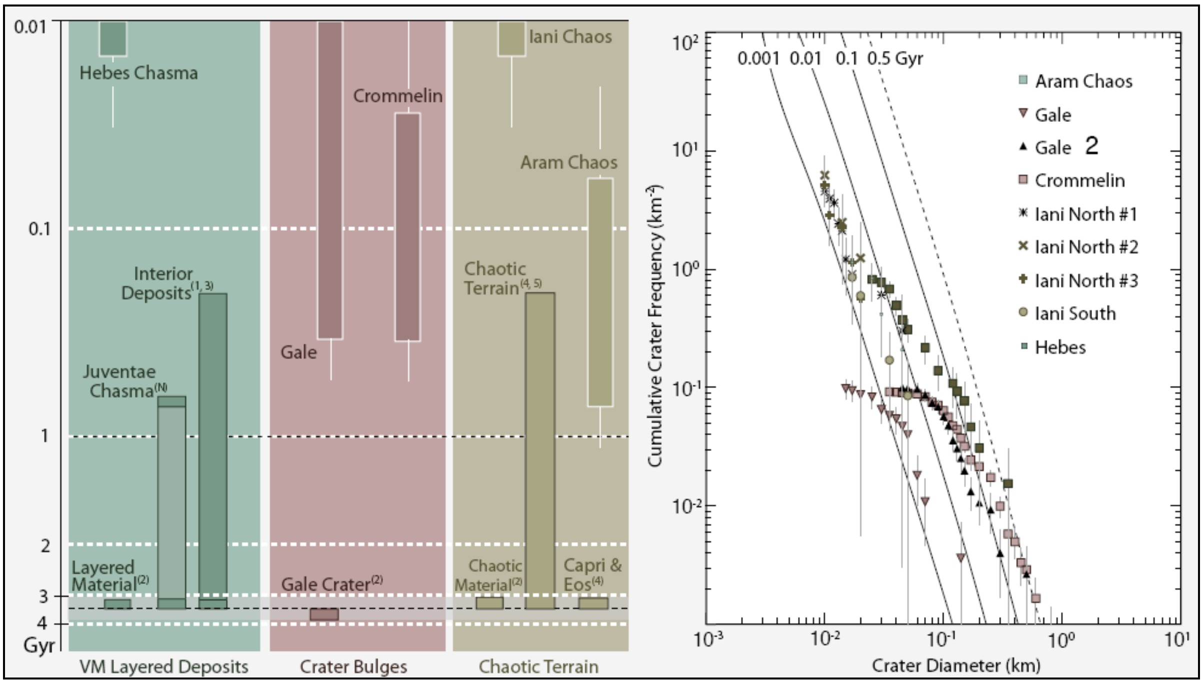

Figure 11: Stratigraphic context (left) for ILDs in Valles Marineris, crater bulges, and chaotic terrain and crater-size frequency distribution (right) as estimated by *Rossi et al.* (2008). Stratigraphic ranges with dark outliners were done by *Neukum et al.* (2007, N), (1) *Lucchitta et al.* (1992), (2) *Scott and Tanaka* (1986), (3) Lucchitta (1999), (4) *Witbeck et al.* (1991), (5) *Rotto and Tanaka* (1995). Methodology and errors are shown in *Neukum and Hiller* (1981). Here, Iani (i.e. Iani 1 and Iani 3) and Aram Chaos and Capri/ Eos Chasma are located in the research area described within this thesis.

2.5 Terrestrial analogues

ILDs display certain morphologies that are similar to terrestrial ones. Mesa and butte morphologies indicate the presence of a former plateau and are thus indicators for heavy erosion by wind, water or ice [*Bahlburg and Breitkreuz*, 1998]. These morphologies are also present in some ILDs. On Earth, these morphologies are known especially from the United States (Fig. 12). Steep and resistant rocks are hardly affected, while soft materials are eroded leaving a mesa and butte morphology [*Bahlburg and Breitkreuz*, 1998]. Buttes look like mesas but are smaller in area and uniform in their dimensions.
These landforms are often composed of sedimentary rock material and occur in arid to semiarid regions on Earth. They are stratified into shale and siltstone and formed by deposition in a meandering system. The weathered siltstone is red due to the occurrence of iron oxide. The top mostly consists of a protective cap rock, which coincides with observations of ILDs on Mars.

Figure 12: East and West Mitten Buttes exposed at the Monument Valley National Park in Utah[1].

Terrestrial tuyas can be compared to ILDs due to their morphology and volcano-tectonic setting [*Chapman and Tanaka*, 2001; 2002; *Chapman et al.*, 2003; *Komatsu et al.*, 2004; *Chapman*, 2007]. Tuyas are subglacial volcanic landforms featuring flat tops and steep-sided flanks (Fig. 13). They are created by lava eruptions into a subglacial water reservoir, which expands and fills with pillows, hyalotuff, hyaloclastites, conglomerates and compound lavas. The size and shape of the tuyas thus formed reflect the reservoir size [*Benn and Evans*, 1998]. They consist of nearly horizontal beds of basaltic lava that caps outward-dipping beds of fragmental volcanic rocks. Tuyas, which are often called Table Mountains (mesas), are common in Iceland.

Figure 13: Hjorleifshofdi Tuya in Iceland (figure capture by U.S. Geol. Surv./ M.G. Chapman). A tuya is characterised by flat top and steep flanks and formed by subglacial volcanism.

Alike, *Chan et al.* (2008) found that the knobs of Candor Chasma on Mars are similar to hoodoos observed in Utah (Fig. 14).
Hoodoos are rock columns made of limestone formed by wind. In general, they are narrow rock needles that form on arid desert ground. They range from 1.5 m up to 45 m in height, whereas Martian knobs are ~30 m high. Their most typical feature is their

[1] http://pdphoto.org

profile formed by wind.

Figure 14: Hoodoos row at the Bryce Canyon National Park in Utah. These morphologies are characterised by their wind-formed profile[2].

Mars features not only these morphologies but also the relevant climatic and depositional conditions. It is important to emphasise that morphological analogues that are best analysed since they are known from Earth, provide hints at Martian scenarios that may account for the formation of ILDs.

[2] http://pdphoto.org