3. Introduction and Motivation

Open Scientific Issues: The current understanding of Mars's geologic history in terms of sequence of events is mainly based on crater size–frequency measurements carried out on high resolution Viking imagery. Prerequisite to the interpretation of crater size–frequency data obtained in various geologic units of different ages is (1) the determination of the shape of the Martian crater production function, implying the primary source of projectiles which impacted the Martian surface, and (2) the application of a reliable cratering chronology model. It has been shown that asteroids from the main belt provided the primary source of impactors on the terrestrial planets in the inner solar system, as inferred from the complex shape of both the crater production function measured on these bodies and the asteroidal size distribution (Neukum, 1983; Neukum and Ivanov, 1994; Neukum *et al.*, 2001; Ivanov *et al.*, 1999; Ivanov, 2001; Werner *et al.*, 2002). For understanding the geologic evolution of a terrestrial planetary body it is necessary to place the different geological processes involved in shaping the planetary surface into a chronological sequence of events. At a regional or local scale, a relative stratigraphy can be derived by analyzing superposition relations and differences in the state of degradation between different geomorphological surface units. Global stratigraphic schemes for planetary bodies are based on the most common resurfacing process: the impacts of planetesimals which remain as crater or crater–related features on planetary surfaces. Through this random cratering process, the counting of the accumulated number of impact craters on planetary surfaces offers a valuable procedure in understanding the chronostratigraphy of a certain object.

Mars has a very diverse impact cratering record in terms of crater morphology, modification and crater frequency. Following the argumentation by Hartmann and Neukum (2001): Various stratigraphic units have been mapped on Mars and their relative ages have been determined by a combination of superposition relations and crater frequencies (Neukum and Hiller, 1981; Tanaka, 1986; Scott and Tanaka, 1986; Greeley and Guest, 1987; Tanaka and Scott, 1987). In principle, absolute ages can be estimated through impact crater frequencies, as it has been shown for the moon. However, the absolute chronology and absolute ages of different Martian stratigraphic units have been known only crudely due to the uncertainties primarily in the Martian impact flux. One approach to extract the crater production function from the geologically distorted record is to transfer the well known and measured production function of the moon to Martian impact conditions considering impact rate and scaling laws (Neukum and Wise, 1976; Neukum *et al.*, 2001; Hartmann and Neukum, 2001; Ivanov, 2001). Until now, due to limited coverage and resolution of the available imagery, it was not possible to measure the Martian crater production function over a wide–enough crater size range. A model for understanding the flux in Martian surroundings has been described by Wetherill (1967, 1979) and has been developed further with respect to dynamical relationships between planet–crossing and main–belt asteroids by Greenberg and Nolan (1989, 1993). Viking and Mariner 9 data analysis, however, led to a wide range of chronologic systems with no clear consensus on the absolute ages (Hartmann, 1973b; Soderblom *et al.*, 1974; Neukum and Wise, 1976; Hartmann *et al.*, 1981; Neukum and Hiller, 1981; Neukum, 1983; Strom *et al.*, 1992).

The most important step, the latest approach by Neukum *et al.* (2001); Hartmann and Neukum (2001); Ivanov (2001) who unified the two competing chronology models (Hartmann, 1973b, 1978; Hartmann *et al.*, 1981; Neukum and Hiller, 1981; Neukum, 1983) and evaluating the two differing styles of crater production functions which appear to agree over most of the diameter range.

Outlined already by Hartmann and Neukum (2001), the earliest Mariner data from 1965 to 1971, revealed heavily cratered terrain where the largest craters (D > 64 km) had crater frequencies similar to those in the lunar highlands, which indicated ages of 3800 to 4500 Ma (Leighton *et al.*, 1965). In the same region smaller craters (250 m <D<16 km) have lower numbers than in the lunar highlands and a wide range of degradation states suggesting losses of smaller craters by erosion and deposition (Öpik, 1965, 1966). This paucity is probably primarily the result of obliteration of craters by erosion and deposition by aeolian, fluvial, glacial and volcanic processes. It is still debated which of these processes is the most important. Although steady rates of obliteration are not applicable, changes in atmosphere thickness (Sagan *et al.*, 1973; Pollack *et al.*, 1987) or increased volcanism (Greeley and Spudis, 1981) or the abundance of permafrost associated with creep deformation (Squyres and Carr, 1986) as well as sporadic glacial erosion are reasonable. Mars Global Surveyor (MGS) added a new twist to understanding the youngest volcanic units by means of higher–resolution images down to 1.5 m/pxl. Crater statistics indicate for restricted areas, e.g. Elysium Planitia, ages less than 100 Ma or even 10 Ma (Hartmann and Berman, 2000). Furthermore, massive layering and mobility of dust and fine material is confirmed by Malin (1998) from MGS images. Investigations of the radiometric crystallization ages of Martian meteorites (SNC) appear to represent mafic igneous intrusions 1300 Ma ago (Nakhlites, Chassigny) and basaltic Shergottites indicate basaltic lava flows about 165 to 475 Ma ago (Nyquist *et al.*, 2001). ALHA 84001 with a crystallization age of 4500 Ma, probably samples the primordial crust of Mars. These meteorites indicate not only the young volcanic activity but also show evidence of liquid water due to alteration products (Shih *et al.*, 1998; Swindle *et al.*, 2000; Sawyer *et al.*, 2000; Bridges and Grady, 2000).

Motivation: One aim of this thesis is to improve and/or verify the existing chronostratigraphic system of Mars. The second goal is to globally understand the geologic evolutionary history of Mars focusing on the volcanic and fluvial processes, giving consistent absolute ages. This implies the photogeologic analysis of all available types of Martian imagery in order to cover all crater diameter ranges to verify the shape of the Martian crater production function. Having been operational at Mars since early 2004, the High Resolution Stereo Camera (HRSC) experiment onboard the European spacecraft MarsExpress introduced the opportunity to gather large image coverage at high resolution (12.5 m/pxl), and allowed measuring crater distributions on various geologic units of different ages. Complemented by data sets collected during the Viking, Mars Global Surveyor, and Mars Odyssey missions at different crater size ranges, the HRSC imagery allows us to determine the "real" shape of the Martian crater production function, not measurable on the previous imagery until now, and to confirm the stability of the crater–generating projectile population for the Martian case. The study also includes detailed investigation of the resurfacing history of the investigated areas and to examine erosion and crater obliteration processes. In parallel, the theoretical treatment, in cases where resurfacing may have occurred, has been developed further and the contribution of background secondary cratering has been investigated in detail to achieve a confident crater size–frequency/age relation.

Structure of the thesis: Firstly, the state of the art regarding the chronostratigraphy, the cratering chronology and the geologic history of Mars is described. Aspects of impact cratering on Mars are outlined in detail (Part I).

In Part II, age dating techniques are introduced. Improved methods for determining absolute ages in cases where resurfacing occurs and the relevance of background secondary cratering are quantified. Finally, the shape of the Martian crater production function for the entire measurable crater diameter range (10 meters to 500 kilometers) is given, based on measurements performed in this thesis. The confirmation and stability in time of the Martian crater production function is essential for the chronostratigraphical investigation.

Applying the knowledge summarized in Part I and II, the Martian stratigraphy is re-assessed (Part III): The determination of the Martian basin formation ages together with the correlation of Martian meteorite crystallization ages and large volcanic units supports the credibility of the Martian cratering chronology model, as it has been transferred from the moon to Mars by Hartmann and Neukum (2001). Detailed investigations of the evolution of the Northern Lowlands, the dichotomy boundary and related fluvial activity (modifying regionally the dichotomy boundary) as well as the global volcanic evolutionary history (in time and space) and the interplay of various processes (volcanic, fluvial, glacial, tectonic and aeolian) have led to a solid data base to finally derive the evolutionary history of Mars with respect to the individual processes.

These resulting global evolutionary aspects are summarized in Part IV, in comparison with previous interpretations of the global chronostratigraphic scheme as it was developed in the post–Mariner and post–Viking era.

With the data gathered in the context of this thesis, an attempt has been made to describe the internal and surface evolution of Mars throughout its whole history.

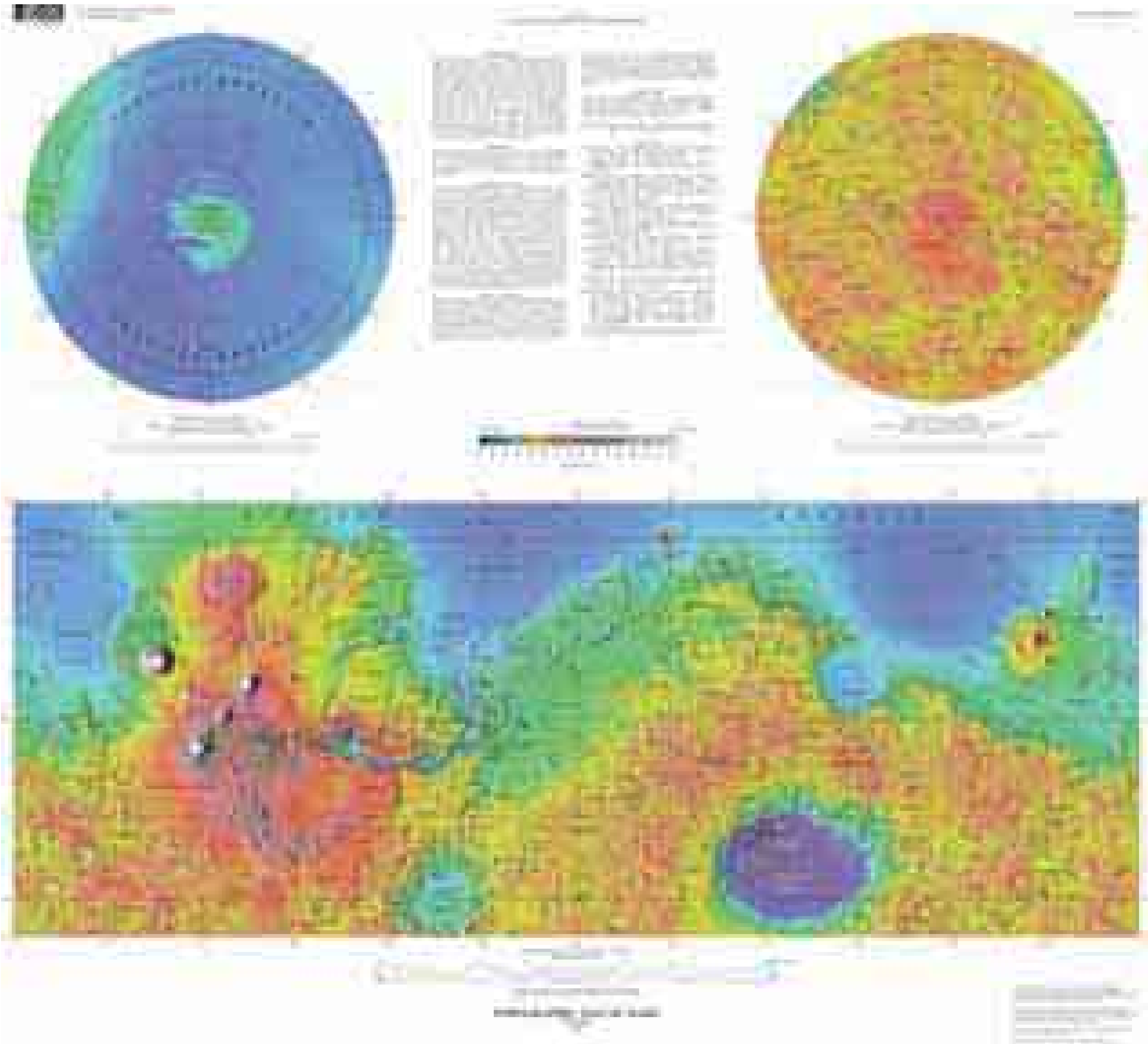

Figure 3.1.: The topographic map of Mars, prepared by the U.S. Geological Survey, based on Mars Orbiter Laser Altimeter data, shall give an overview of locations and general surface features as they are discussed in this thesis.

Part I.

Introduction to the Background Theory and Open Scientific Issues

4. The Base of Our Knowledge – The Moon, Earth and Venus

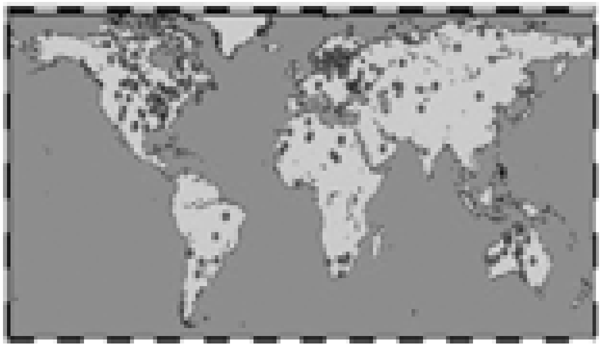

Figure 4.1.: Impact crater Distribution on Earth: Due to plate tectonics and erosional processes the impact crater population on Earth has been strongly modified. Especially oceanic crust is widely unaffected by craters due to its young age (less than 175 Ma). (Source: Earth Impact Database at the PASSC, University of New Brunswick, Canada)

The Moon, our closest neighbour, is the best studied extraterrestrial object in our solar system. Moreover, the target of the first space missions culminating in manned landing and sample–return during the American Apollo and Russian Luna Programs (1969–1976). Therefore, the Moon is the only planetary body, apart from Earth, where we can relate rock samples as well as their composition and age, more or less confidently to specific geomorphologic units. Furthermore, based on isotope ratios and chemical composition, the differentiation of the Moon ended long ago, and based on its small size the Moon is considered to be no longer geologically active. The two dominant geological processes that have sculpted the lunar surface are impact cratering and mare volcanism. Volcanism as a surface shaping process has ended but impact cratering is still ongoing and is a process relevant to this thesis. The Moon has preserved much of its surficial magmatic and cratering record for most of its life span. Lunar surface interpretation allows us to look back into the early phase of planetary evolution, while on Earth, plate tectonics and resurfacing processes driven by large–scale mantle convection, change the visible surface and have erased most of its impact record. The current impact crater distribution on Earth (Fig. 4.1) is basically controlled as on any other extraterrestrial solid–surface body by its surface or crustal age.

The Earth

The densely cratered units are correlated with the cratons (based on crater counts, their mean age is $\sim$ 400 Ma, (Neukum, 1983), while the crystallization ages of the rock could vary). The lack of craters in the oceanic crust is due to two reasons: (1) In general, oceanic crust is very young (on average 62 Ma; maximum age is $\sim$ 175 Ma) and almost no craters are expected (Koulouris *et al.*, 1999). (2) Small impactors (less than $1 - 2$ km in diameter) do not affect the ocean floor because the motion of such projectiles is already decelerated when they reach a depth of about 1.5 km (Shuvalov, 2003) while the average depth of the ocean is about 5 km. Only a single impact event in deep water is known and there no crater is present: Eltanin in the Bellingshausen Sea (Gersonde *et al.*, 1997). Crater records on land of the smaller–size range ($<$ 5 km) can be either erased by geological processes, e. g. erosion, or later exhumed after being buried under a protecting cover. Hence their frequency might not represent the real surface age (e.g. in Fennoscandia). Both processes have to be considered when crater size–frequency distributions are measured on planetary surfaces which are geologically more active than the Moon.

Venus

An additional factor on terrestrial crater size–frequency distributions is the shielding effect of the atmosphere, which on Venus is even stronger than on Earth (Ivanov *et al.*, 1992; McKinnon *et al.*, 1997). On Venus projectiles producing craters of diameters less than approximately 1 km are disrupted and possibly hit the surface as a swarm of fragments creating impact crater clusters, if they reach the surface at all. On the other hand, the recognition of craters on Earth's surface is highly biased by the accessibility of the area and the research activity within a specific region.

The Earth can be used as an analogue, but has its clearly limitations when exploring the past, although the well-dated rock record is of tremendous value. Therefore, the Moon has become, based on its cratering record, a time–stratigraphic calibration reference for the Earth–Moon system and the entire inner Solar System. A combination of interpreting ground–based and orbiter–based multispectral imagery and in–situ data promote the Moon (apart from the Earth) to be a valuable test case and calibrator for many methods based on remote sensing data.

4.1. Geologic Evolution of the Moon

The formation and evolution of the Moon is part of the evolution of the planetary system. Even in modern times, hypotheses of the origin of the Moon include capture from an independent solar orbit, rapid co–accretion or fission from the Earth. A variant of the fission hypothesis is that the Moon accreted rapidly from ejecta of a massive (proto–mars sized) impact on the Earth after its core had formed (Hartmann and Davis, 1975; Cameron and Ward, 1976; Cameron, 1986). These later alternatives are based on the commonly accepted idea that the Moon accreted from material similar to the Earth, as clearly shown by oxygen isotopes (Clayton and Mayeda, 1975; Hartmann *et al.*, 1986; Halliday, 2000). It is required that the Moon initially was in a completely or partially molten state (Canup and Agnor, 2000). Tidal interaction between the Earth and the Moon led to a quite stable orbital configuration of the satellite having a spin period equal to its orbital period, but slowing down the Earth with the Moon receding from the Earth (opposite to the Mars – Phobos situation, where the satellite is believed to move towards the main body (Yoder, 1995)). Due to the synchronous rotation of the Moon only half of the lunar surface is visible from Earth (near–side). The Moon, lacking an atmosphere, allows us even with the naked eye to identify bright and dark regions: the Face of the Moon. The near–side of the Moon is characterized by extensive darkish almost flat lava plains (Maria), mountain chains and brighter heavily cratered highland plateaus (Fig. 4.2).

The brighter plateaus are of anorthositic composition, and the primordial crust originated in the very beginning of the lunar geological history about 4.6 to 4.5 Ga ago (Halliday, 2000). This composition suggests fractional crystallization and differentiation from a global magma ocean (Taylor, 1982). The second stage of the evolution of lunar highland crust was its modification during a period from 4.4 to 3.9 Ga through the crystallization of mafic and ultramafic plutonic rocks (Shearer and Papike, 1999; James, 1980). Because the Moon, as all terrestrial planets, experienced a period of heavy bombardment (until 3.9 Ga ago), all of these early periods of magmatism, crust formation, and lunar mantle evolution are not preserved. The early upper lunar crust is perforated by impact craters of diverse sizes, covered by their polymict ejecta, and fractured and brecciated to depths of up to 20 km (megaregolith, Hartmann, 1973a). Mare basaltic magmatism followed the heavy bombardment period and basin formation, and occurred between 3.9 and 2 Ga ago. In small areas the gradual decay of volcanic activity even went on until 1 Ga ago, (e. g. Nyquist and Shih, 1992, and based on crater counts: Schaber (1973), or Hiesinger *et al.* (2002)).

Whereas impact cratering is a spatially random process, the maria are far from uniformly distributed. Images from the Moon's farside (Fig. 4.2) clearly show that more than 90% of the surface area of the lunar basaltic lava flows are located on the Earth–facing hemisphere,