Ion Escape from Mars: Measurements in the Present to Understand the Past

-Doctoral Thesis-

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"Ah! Don’t say you agree with me. When people agree with me I always feel I must be wrong."

– Oscar Wilde (The Critic as Artist, 1891)
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Abstract

Present-day Mars is a cold and dry planet with a thin CO$_2$-dominated atmosphere comprising only a few mbar pressure at low altitudes. However, the Martian surface is marked with valley networks, hydrated mineral clays, carbonates and the remains of deltas and meandering rivers, i.e. traces of an active hydrological cycle present early in the planet’s geological history. A strong greenhouse effect, and thus a thicker atmosphere, would have been required to sustain a sufficiently warm environment, particularly under the weaker luminosity of the early Sun. The fate of this early atmosphere is currently unknown.

While several mechanisms can remove atmospheric mass over time, a prominent hypothesis suggests that the lack of an intrinsic Earth-like global magnetic dipole has allowed the solar wind to erode the early Martian atmosphere by imparting energy to the planet’s ionosphere which subsequently flows out as ion escape, over time depleting the greenhouse gasses and collapsing the ancient hydrological cycle. Previous studies have found insignificant ion escape rates under present-day conditions, however, the young Sun emitted significantly stronger solar wind and photoionizing radiation flux compared to the present. The geological record establishes the time of collapse of the hydrological cycle, estimated to have occurred in the mid-late Hesperian period (\(\sim 3.3\) billion years ago) at latest. To constrain the amount of atmosphere lost through ion escape since, we use the extensive database of ion flux measurements from the Analyzer of Space Plasmas and Energetic Atoms (ASPERA-3) particles package on the Mars Express orbiter (2004-present) to quantify the ion escape rate dependence on upstream solar wind and solar radiation conditions.

The Martian ion escape rate is shown to be insensitive to solar wind parameters with a weak inverse dependence on solar wind dynamic pressure, and linearly dependent on solar ionizing photon flux, indicating efficient screening of the bulk ionosphere by the induced magnetic fields. The impact of an extreme coronal mass ejection is studied and found to have no significant effect on the ion escape rate. Instead, intense solar wind is shown to only increase the escaping energy flux, i.e. total power of escaping ions, without increasing the rate by accelerating already escaping ions. The orientation of the strongest magnetized crustal fields are shown to modulate the ion escape rate, though to have no significant time-averaged effect. We also study the influence of solar wind and solar radiation on the major Martian plasma boundaries and discuss factors that might limit the ion escape rate, including solar wind–ion escape coupling, which is found to be \(\lesssim 1\%\) and decreasing with increased solar wind dynamic pressure. The significant escape rate dependencies found are extrapolated back in time, considering the evolution of solar wind and ionizing radiation, and shown to account for only \(4.8 \pm 1.1\) mbar equivalent surface pressure loss since the time of collapse of the Martian hydrosphere in the Hesperian, with \(\sim 6\) mbar as an upper estimate. Extended to the late Noachian period (3.9 billion years ago), the found dependencies can only account for \(\lesssim 10\) mbar removed through ion escape, an insignificant amount compared to the \(\gtrsim 1\) bar surface pressure required to sustain a warm climate on early Mars.
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Sammanfattning


Flera mekanismer kan dock minska atmosfärrens massa över tid. En vanlig hypotes är att frånvaron av en jordliknande magnetisk dipol har låtit solvinden erodera Mars atmosfär genom att överföra energi till partiklarna i planetens jonosfär som sedan flyr ut i rymden. Över tid töms atmosfären på växthusgaser och vattencykeln kollapsar. Tidigare studier har funnit att jonutflooten är insignifikant, men solvinden och den joniserande solstrålningen var mycket starkare i det tidiga solsystemet än dom är idag och kan ha drivit högre flöden. Mars geologiska historia etablerar när dess hydrologiska cykel stannade, vilket uppskattas ha skett för 3.3 miljarder år sedan som senast, under den hesperiska perioden. Vi begränsar uppskattningen av den totala atmosfärmassa som flytt som joner sedan dess genom att kvantifiera jonutflooten samt dess beroende på solvinden och den joniserande solstrålningens egenskaper med partikelinstrumentpaketet Analyzer of Space Plasmas and Energetic Atoms (ASPERA-3) på Mars Express.

Vi visar att jonutflooten är okänsligt för solvinden, med ett svagt invers beroende på solvindens dynamiska tryck och ett lineärt beroende på solens flöde av joniserande fotoner, detta indikerar att de inducerade magnetfälten väl avskärmar huvuddelen av jonosfären från solvinden. Vi studerar en extrem koronamassutkastning som träffade Mars och finner likaså att denna inte ökade jonutflooten. Intensiv solvind visas istället endast öka energiutflödet genom att accelerera redan flyende joner. Riktening hos de ytor med starkt magnetiserad skorpa visas modulera jonutflooten men även att inte ha någon signifikant total effekt på jonutflooten över tid. Vi studerar även solvindens påverkan på de främsta plasmaområdetsgränserna och diskuterar faktorer som kan begränsa jonutflooten, inklusive koppling mellan solvinden och jonosfären, vilken finns vara ≲1% och minskar med solvindens dynamiska tryck. De funna signifikanta förhållandena extrapoleras bakåt i tiden, beaktandes solvindens och den joniserande solstrålningens evolution, vilka visas ge upphov till en förändring i atmosfärmassa motsvarande 4.8 ± 1.1 mbar sedan vattencykeln kollapsade under den hesperiska perioden. Som en högre uppskattning ges ≈6 mbar. Utökat till den sena noakiska perioden (3.9 miljarder år sedan) beräknas ≲10 mbar ha flytt som joner, en insignifikant mängd i jämförelse med ≳1 bar som behövs för att uppehålla ett varmt klimat under Mars tidiga geologiska historia.
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List of appended papers


List of related papers


Continued on next page


Acknowledgements

Over the course of my studies at the Swedish Institute of Space Physics/Institutet för rymdfysik (IRF) I have had the privilege to work with the many dedicated researchers and staff here. This thesis and the body of work behind it would not have been possible to complete without the supportive people and environment that make IRF a fantastic place to do science.

I am deeply grateful to my main supervisor and mentor, Stas Barabash; for offering the opportunity to work on ion escape and much freedom in this work, for the many ideas sparked on late nights attempting to unravel the fundamentals of atmospheric escape, for showing me not only how to do good science, but also how to be a good scientist.

I would like to thank my assistant supervisor Yoshifumi Futaana, I am sincerely grateful for all the advice and insights he has shared over the years, always with a smile. Many thanks to Hans Nilsson for invaluable help with understanding and analyzing the Mars Express data, and to Mats Holmström especially for providing the means to complete my final year.

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Preface

The ultimate goal of the work behind this thesis is to establish the factors that influence the atmospheric ion escape rate and thus provide an estimate for the total amount of atmosphere lost through ion escape since the collapse of the early Martian hydrosphere. While independently the work is firmly the subject of space physics, when and how the Martian hydrosphere collapsed are fundamental planetological problems. Thus, a significant portion of the first chapter is dedicated to introduce the reader to Martian geological history as it serves as a large part of the rationale for studying the atmospheric escape rate. The following sections and chapters aim to provide the reader with a superficial overview of escape processes, the solar cycle, the Martian plasma environment, the relevant instrumentation, and to give a summary of the method and results of the appended papers. However, the reader should be aware that Mars is an object of intense study, perhaps now more than ever, with several missions typically arriving with every available launch window, adding to our knowledge of the planet.

During the course of my studies at IRF I have also had the pleasure to contribute to, and be part of, several related projects spanning Martian ionospheric studies, space weather, the plasma environment of comet 67P/Churyumov–Gerasimenko, and ion escape from Venus and Earth. The associated results that are published or in publication at the time of writing are listed as "Related papers" on page vii.

Descriptions and credits for header images as follows. Contents: Mars Express in orbit of Mars, watercolor and pencil on paper. Ch. 1: Panoramic photograph of Yellowknife Bay, an ancient lake- and streambed in Gale crater, Mars. Taken by NASA's Curiosity rover at twilight. Credit NASA/JPL. Ch. 2: Mars, the Sun and the solar wind, based in part on an illustration by [Brain et al., 2017], watercolor on paper. Ch. 3: 3D model of Mars Express, Mars in background. Ch. 4: View of Argyre basin from orbit, composition of photographs taken by the NASA Viking orbiter, credit NASA/JPL/Daniel Macháček. Bibliography: Crater on the horizon, watercolor and pencil on paper. Headers for Contents, Ch. 2, and Ch. 5 courtesy of Anastasia Grigoryeva.

Leif Eric Robin Ramstad
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Abstract ................................................................. i
Sammanfattning .......................................................... iii
List of appended papers .................................................... v
List of related papers ....................................................... vii
Acknowledgements ........................................................ ix
Preface .................................................................... xi

1 Introduction ............................................................... 1
1.1 The present and early planet Mars ................................. 1
1.1.1 Noachian period .................................................. 4
1.1.2 Hesperian period .................................................. 6
1.1.3 Amazonian period ............................................... 7
1.1.4 Climate modelling and interpretation of the geological record ........................................... 9
1.2 Mechanisms for atmospheric loss .................................. 10
1.2.1 Ground sequestration .......................................... 10
1.2.2 Escape to space .................................................. 10

2 Solar wind interaction with Mars ..................................... 17
2.1 The Sun, solar wind and interplanetary magnetic field ........ 17
2.1.1 Properties of the solar wind .................................... 17
2.1.2 Solar radiation ................................................... 18
1. Introduction

1.1 The present and early planet Mars

Mars is a small terrestrial planet in the inner solar system, only about half the diameter of the Earth and barely 1/10 its mass. The tenuous, CO$_2$-dominated atmosphere averages only 6.4 mbar at mean elevations, creating a modest ∼5 K greenhouse effect with negligible influence on the cold dry surface environment, as well as large diurnal variations in local surface temperatures, as these are thus primarily controlled by solar irradiance. By extension, the high-latitudes experience extreme polar nights where temperatures drop sufficiently for the atmosphere to deposit directly onto the ground, seasonally covering the poles in a layer of CO$_2$ ice up to 2 m thick [Smith, 2001], modulating the global atmospheric pressure by ∼25%. A global north-south hemispheric elevation dichotomy separates the relatively young northern low-lying plains from the generally older southern highlands (see Figure 1.1) and broadens the distribution of surface pressures. The relatively large 0.093 eccentricity of the Martian heliocentric orbit modulates the global solar input by up to 46%, further widening the range of conditions.

Liquid water is generally not stable under present-day Martian surface temperatures and pressures. However, there is evidence that the diversity in surface conditions allows for localized, seasonal appearances of liquid perchlorate brine on present-day Mars. Recurring slope lineae appear as incrementally growing low-albedo streaks on steep and relatively warm equator-facing slopes between

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
</tr>
</thead>
<tbody>
<tr>
<td>Sun distance</td>
<td>1.38 (1.52) 1.67 AU</td>
</tr>
<tr>
<td>Orb. period</td>
<td>1.88 Earth-years</td>
</tr>
<tr>
<td>Orb. ecc.</td>
<td>0.093</td>
</tr>
<tr>
<td>Rot. period</td>
<td>24 h 37 m 22 s</td>
</tr>
<tr>
<td>Obliquity</td>
<td>25.19°</td>
</tr>
<tr>
<td>Mean radius</td>
<td>3390 km / 0.53 R$_⊕$</td>
</tr>
<tr>
<td>Mass</td>
<td>0.107 M$_⊕$</td>
</tr>
<tr>
<td>Surf. gravity</td>
<td>0.376 g</td>
</tr>
<tr>
<td>Surf. pressure</td>
<td>5.6 (6.4) 7.2 mbar</td>
</tr>
<tr>
<td>Surf. temp</td>
<td>-143 (-63) +20 °C</td>
</tr>
</tbody>
</table>

Table 1.1: Basic Mars planetary and orbital parameters in the present day for reference. Ranges indicate variations estimated close to average surface elevation with mean values in parenthesis.
Figure 1.1: Mars global topography map derived from Mars Global Surveyor/Mars Orbital Laser Altimeter (MGS/MOLA) data. The color scale is divided above and below the areoid. Accordingly, blue colors mark lowlands and red colors mark highlands, showing the global dichotomy of the Martian surface. Lines show locations of mapped valley networks [Hynek et al., 2010], i.e. erosion features left by precipitation, which are primarily found in the generally older highlands. The geographical grid is equally spaced 30° in latitude and longitude, centered on the prime meridian.

Kasei Valles is one of the largest and visually most striking examples of Martian fluvial geomorphology, a large outflow system situated north of the recognizable Valles Marineris canyon, both visible in Figure 1.1 (the mouth of Kasei Valles at ca. 30°N, 60°W). Its detailed topography is pictured in Figure 1.2, clearly showing the patterns left by flows in the past. Kasei Valles illustrates well the difficulty in deciphering the ancient Martian climate. There is evidence of at least four distinct formative intervals covering all the Martian time periods [Chapman et al., 2010], during which the source area likely changed due to the rise of the volcanic Tharsis plateau to the west and features left by the oldest flows were partially removed by newer flows as well as buried by lava.
1.1 The present and early planet Mars

Figure 1.2: Kasei Valles, one of the largest and visually most striking Martian outflow systems, draining into the Chryse Planitia plains to the east, likely an ancient impact basin. The source region of the remnant system is Echus Chasma, near the equator. However, Kasei Valles has formed over time through all major time periods [Chapman et al., 2010], during which the source region has likely changed several times due to the emergence of the Tharsis plateau immediately to the west, illustrating the difficulties in deciphering the Martian geological history. Altimetry is derived from MGS/MOLA data.

The Martian geological record is commonly divided into three major periods based on crater impact distribution, here defined in time before the present; the Noachian (4.1–3.7 Ga), the Hesperian (3.7–3.0 Ga), and the Amazonian (3.0 Ga to the present). The term Pre-Noachian is tentatively used to refer to the period from formation to 4.1 Ga, however no Pre-Noachian terrains have been conclusively indentified, thus very little is known about this period [Mangold et al., 2016]. The three periods mark time-spans that are qualitatively different in geomorphological and mineralogical surface records, though the exact transition times are not accurately established. It is clear that large volumes of water were moved from low to high elevations to allow the fluvial processes that created the many landforms on the surface, implying the existence and subsequent collapse of an ancient evaporation-precipitation cycle, i.e. an active hydrosphere, on early Mars. Under the weaker luminosity of the early Sun, a thick atmosphere would have been required for a strong greenhouse effect to maintain temperatures above 0 °C at low altitudes. It is currently not clear exactly when or why the Martian hydrosphere collapsed, however the geological records from the individual time periods provide some constraints.
1.1.1 Noachian period

Noachian (4.1–3.7 Ga) terrains are in part characterized by the presence of well-developed valley networks (see Figure 1.1), e.g. the prominent Parana Valles and tributaries to downstream Loire Valles [Bouley et al., 2010]. Valley networks are thought to form from persistent precipitation over a widespread area, where the calm drainage of water over time erodes a dendritic system of valleys. Hypothetically, valley networks can also form from large-scale groundwater seepage, however, the shape and distribution of Martian valley networks are consistent with precipitation [Craddock and Howard, 2002]. A prominent example of Noachian valley networks near Huygens crater is shown in Figure 1.3, based on Mars Express High Resolution Stereo Camera (MEX/HRSC) and MGS/MOLA data.

![Figure 1.3](image)

**Figure 1.3**: Well-developed Noachian-period valley networks on the outer flank of Huygens crater (bottom left depression) that were draining into and away from Huygens. Several networks were draining into the center crater with a smoothed floor, such smooth craters are in many cases thought to be filled with sedimentary deposits laid down in ancient lacustrine environments. The presence of well-developed valley networks in Noachian terrains indicate persistent precipitation over long time spans [Craddock and Howard, 2002; Fassett and Head, 2008]. Note that the eastern low-lying area is Hesperian in age. Composite of MEX/HRSC data product h0532_0000.nd3.05.jp2 and MGS/MOLA global altitude data for pixel intensity and color, respectively.
1.1 The present and early planet Mars

While the presence of valley networks are strong indicators for long-term precipitation, the presence of stratified Fe/Mg-dominated phyllosilicate clay mineral deposits in Noachian terrains also indicates that iron- and magnesium-bearing minerals were weathered by persistent contact with neutral or alkaline aqueous solutions for extended periods of time [Poulet et al., 2005; Bishop et al., 2008; Ehlmann et al., 2009]. Clay deposits are also commonly found in association with carbonate minerals [Ehlmann et al., 2008], which form from another aqueous alteration process discussed later in this chapter, strongly indicating persistent lacustrine environments as the common formative medium. Two locations on the Martian surface are particularly rich in phyllosilicate clay deposits; the hills surrounding Mawrth Vallis (ca. 24°N, 28°W) and the grabens of Nili Fossae (ca 22°N, 76°E), both Noachian in age. A high-resolution view of deposits in Mawrth Vallis is shown in Figure 1.4a.

Figure 1.4: Evidence for aqueous environments on early Mars, on a wide variety of scales. a) Layered deposits rich in phyllosilicate clays on the rim separating two impact craters in the western hills surrounding Mawrth Vallis, as observed by the High Resolution Imaging Experiment on Mars Reconnaissance Orbiter (MRO/HiRISE). The spatial range of clay deposits near Mawrth Vallis extend to nearly 300 × 300 km. b) Cross-bedded strata formed by deposition in a fluvial environment observed by Mars Science Laboratory in Gale crater, which formed near the Noachian-Hesperian boundary. c) Inverted remains of a delta in Eberswalde crater observed by the Context Camera (CTX) on MRO.
1.1.2 Hesperian period

The Hesperian (3.7–3.0 Ga) is commonly regarded as a transition period between the warm and wet Noachian environment and the dry, cold Amazonian. However, the exact time of this transition is not well constrained and has been further complicated by studies indicating that the transition may not have been neither continuous nor well-defined by a single point in time. In-situ investigations by the Mars Science Laboratory (MSL) rover have shown that the floor of the ∼145 km diameter wide Gale crater (5.4°S, 137.8°E) is the erosional remnant of sedimentary deposits collected between the crater’s formation 3.8–3.6 Ga and exhumed before 3.3–3.1 Ga [Grotzinger et al., 2015], i.e. mainly in the Hesperian. The current floor exposes sediments of varying ages containing mudstones with significant phyllosilicate compositional components [Vaniman et al., 2014] which formed in a lake that was sufficiently deep to maintain stratified oxidizing and redox environments [Hurowitz et al., 2017]. Gale crater also hosts cross-bedded strata, which form by deposition in fluvial environments (see Figure 1.4b). Geochemistries of mudstone deposits in Gale crater indicate increasing temperatures in the early Hesperian [Hurowitz et al., 2017]. Whether this increase is a fluctuation or a long-term trend will hopefully be revealed in the coming years as MSL climbs to higher, younger strata on Aeolis Mons/Mount Sharp.

The Noachian-Hesperian boundary roughly coincides with a change in global aqueous chemistry from neutral/alkaline solutions to acidic, as observed by a transition from phyllosilicate clay deposits to sulphates (e.g. kieserite MgSO_4\cdot H_2O and gypsum CaSO_4\cdot 2H_2O). The change in aquatic acidity is thought to be the result of a rise in atmospheric SO_2, released by the extensive volcanism in the late Noachian and early Hesperian that formed the Tharsis volcanoes [Bibring et al., 2006].

Many of the large outflow channels were carved from catastrophic floods released during the Hesperian (e.g. large parts of Kasei Valles [Chapman et al., 2010], Figure 1.2). However, the short time-scales of such events provides little climatic information as their occurrence do not necessarily imply stable liquid surface water [Mangold et al., 2016].

Nevertheless, Hesperian terrains do feature morphological signatures of persistent fluvial activity. Clear examples include the inverted, though spectacularly well-preserved, shapes of rivers and deltas in the Aeolis regions (ca 3°S, 147°E). Aeolis planum is part of the Medusae Fossae formation, which is thought to have been deposited pyroclastically in the Hesperian [Kerber and Head, 2010]. Stratified inverted river deposits (see Figure 1.5) in the large sedimentary basin Aeolis Dorsa have recently been interpreted to indicate a moving shoreline indicating large changes in the Aeolis Dorsa paleolake [Cardenas et al., 2017]. Perhaps the youngest strong evidence for precipitation in the Hesperian are the valley networks in Warrego Valles (42°S, 92°W) on the south-eastern flank of the Tharsis bulge, pictured in Figure 1.5. These networks formed between 3.5–3.0 Ga, i.e. in the mid-late Hesperian, and while not as well-developed as their Noachian counterparts (compare Figure 1.3), they might constitute some of the youngest traces of an active hydrological cycle [Ansan and Mangold, 2006], and may thus be regarded as a lower limit for the age of collapse of the hydrosphere.
1.1 The present and early planet Mars

1.1.3 Amazonian period

The Amazonian is the current Martian time period (3.0 Ga to the present), characterized by a cold and dry climate with the remaining surface water trapped in the polar caps, glaciers or permafrost deposits, i.e. the cryosphere.

Precipitation appears to have ceased by the early Amazonian, and fluvial morphology in Amazonian terrains is instead dominated by large outflow events triggered by volcanism. Some of the most spectacular remains are left by the most recent large outflow event, triggered by the opening of the Cerberus Fossae fissures (10°N, 157°E) only about 2.5 Ma ago [Vaucher et al., 2009]. The fissures are thought to have released the water contained in a pressurized deep aquifer onto the surface. The catastrophic flood carved out the Athabasca Valles, leaving streamlined islands behind hardened structures such as craters. A particularly dramatic example of a crater eroded by the Athabasca floods is shown in Figure 1.6.

There are also smaller, more widespread, fluvial features present in some

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**Figure 1.5:** Examples of climatically relevant Hesperian fluvial landforms. Left panel: Inverted remains of a meandering river and possible oxbow lake (dashed curve) in Aeolis Dorsa (6°S, 153°E). Right panel: Valley networks in Warrego Valles, some of the youngest strong evidence for precipitation on Mars, and thus an active hydrosphere in the mid-late Hesperian (3.5–3.0 Ga) [Ansan and Mangold, 2006]. Composite of MEX/HRSC data product h0453_0000_nd3.jp2 and MGS/MOLA altimetry.
Chapter 1. Introduction

Amazonian terrains. However, these are typically associated with localized transient events as these are shallow, less developed and typically located in relation to volcanoes and impact craters which provide heat to melt ice deposits [Mangold et al., 2016]. For example, channels surrounding the ∼1.0 Ga old Hale crater coincide with distribution of hot impact ejecta that are thought to have melted subsurface volatiles [Jones et al., 2011].

Figure 1.6: Height profile of a crater in Athabasca Valles which has been heavily eroded by catastrophic flooding caused by release of subsurface water with the opening of the Cerberus Fossae fissures to the north-east (location context in upper-left panel, crater location marked by red cross). Besides the streamline shape of the wake, the rapid flow of water has left giant current ripples (magnified in upper right panel). The Athabasca floods were geologically very recent events, estimated to have occurred only ∼2.5 Ma ago [Vauchet al., 2009], and typical of large-scale fluvial activity in the cold and dry Amazonian period, i.e. associated with volcanism or local impact heating, rather than climate.
1.1 The present and early planet Mars

1.1.4 Climate modelling and interpretation of the geological record

Multifaceted geomorphological, chemical, orbital and in-situ observations provide ample evidence of an active Martian hydrosphere throughout the Noachian and well into the Hesperian, tentatively with an optimum in the early Hesperian [Ansan and Mangold, 2006; Hurowitz et al., 2017] due to continuous volcanic outgassing gradually supplementing the remaining primordial atmosphere with up to ∼0.8 bar CO$_2$ and SO$_2$ [Craddock and Greely, 2009]. The Martian surface underwent intense aqueous activity for a total time period that can be estimated between several 100’s Ma up to ∼1 Ga [Mangold et al., 2016].

It is currently unclear what properties allowed the ancient Martian atmosphere to maintain surface temperatures well above the freezing point of water for such a long period of time, when the bolometric luminosity of the Noachian/Hesperian Sun was 70-80% that of today. This is the classic problem commonly called the Faint young Sun paradox. Climate modelling efforts have attempted to assess the atmospheric pressure and composition requirements to maintain a warm, wet early Martian climate, finding that a dense CO$_2$-rich atmosphere is required not only for a stable high greenhouse radiative surface balance, but also for efficient equator-pole heat transport to avoid greenhouse collapse by collection of the global water content at the poles [Wordsworth et al., 2015].

Nevertheless, the Martian greenhouse effect and water cycle evidently did collapse, though at a time that remains loosely constrained. The formation of the Warrego Valles valley networks in the mid-late Hesperian could tentatively provide a rough lower limit, however, the rate of erosion in this area is not well constrained and short formation time-scales are expected (1 ka–1 Ma), which could imply formation under episodic warm wet periods in an otherwise potentially cold, dry mid-late Hesperian climate [Ansan and Mangold, 2006]. Radiative transfer models have demonstrated strong warming potentials from inclusion of minor species H$_2$ and CH$_4$ by collision-induced absorption (CIA) in CO$_2$-dominated (0.5-3.0 bar) atmospheres [Ramirez et al., 2014; Ramirez, 2017; Wordsworth et al., 2017]. These species are expected to have been significant constituents in the primordial reducing atmospheres of all the terrestrial planets, however, their true contents and hypothetical periodic releases [Wordsworth et al., 2017] in the evolved Hesperian atmosphere are not well constrained, and neither is the CO$_2$ partial pressure.

The recent study by Bristow et al. [2017] potentially provides an empirical constraint. The authors report concentrations of carbonates below detectable limits in ∼3.5 Ga (Hesperian) mudstones as measured by MSL in Gale crater, implying CO$_2$ pressures on the order 10’s of mbar despite the presence of a lacustrine environment, adding to similar low estimates of the Martian carbonate inventory [Edwards and Ehlmann, 2015].

However, practically all recent climate models require assumptions for CO$_2$ pressures ≳1 bar to explain a stable early Martian greenhouse effect and water cycle. This assumption is not consistent with the current ∼6.4 mbar CO$_2$ inventory nor the Hesperian estimation by Bristow et al. [2017], implying either that 1: CO$_2$ has been rapidly depleted from the atmosphere over time and the estimate from Gale crater is unrepresentative of the global conditions, perhaps due to carbonate weathering, or 2: that an unknown atmospheric or geological component supplied or trapped
sufficient heat to maintain the hydrosphere. We may consider the first hypothesis by constraining the total atmospheric loss over time, however, the dominant processes responsible are currently unknown, in the next section we will introduce and discuss some candidates.

## 1.2 Mechanisms for atmospheric loss

The Martian atmosphere is neighboured by two domains: the surface and outer space. Thus, loss of atmospheric mass implies either sequestration in the ground, or escape to space. The processes involved are fundamentally different, sequestration is driven by chemical reactions while escape to space requires acceleration of atmospheric particles past escape velocity in a non-collisional regime. The following sections will introduce the mechanisms and reactions involved as well as give current available estimates as to their influence on atmospheric loss for Mars.

### 1.2.1 Ground sequestration

Liquid water in contact with atmospheric CO$_2$ forms a weak carbonic acid solution with a temperature- and pressure-dependent carbonate anion concentration [CO$_{3}^{2-}$]. On a planetary surface the liquid water will also be in contact with the various minerals in the surrounding rock, releasing dissolved metal cations [$M^{X+}$] that react with the carbonate anions, forming various species-dependent carbonate minerals.

Magnesite (MgCO$_3$) is the most common carbonate species detected at Mars, likely weathered from olivine ((Mg,Fe)$_2$Si$_2$O$_6$) and observed through its near-infrared (NIR) 2.3 and 2.5 $\mu$m absorption bands. Magnesite is found in the regolith at low concentrations (∼2-5 w%) [Bandfield et al., 2003], and in exposed carbonate deposits [Ehlmann et al., 2008]. The exposed deposits on Mars are dominated in quantity by a single location, the Nili Fossae grabens (23°N, 77°E), which have been estimated to contain carbonates equivalent to only 0.25–12 mbar of dissolved CO$_2$ [Edwards and Ehlmann, 2015], an insignificant amount in comparison to the ∼1 bar pressures required from most climate models.

The apparent lack of Martian carbonate deposits suggests that either most of the pre-Hesperian atmosphere has been lost to space, or the overwhelming majority of carbonate deposits are buried and thus hidden from detection by orbiting NIR spectrometers. A recent study by Wray et al. [2016] reports evidence for buried layers of siderite (FeCO$_3$) and calcite (CaCO$_3$) exposed by eroded valleys and impact craters in the western Noachian highlands. The walls of Her Desher Vallis (25°S, 48°W) shows a particularly clear shallow subsurface layer of Fe/Mg-phyllosilicates mixed with Fe/Ca carbonates, indicating a common aqueous formative environment.

While remnant surface carbonate deposits are small, it is currently unclear how large global quantities of equivalent CO$_2$ could be sequestered in such subsurface layers.

### 1.2.2 Escape to space

The Martian atmosphere is bound to the planet by its gravitational potential field. On a global scale Mars is large enough to be well-described as a sphere, thus we can
describe the gravitational potential outside the planet’s surface using the spherically symmetric Newtonian approximation

\[ V_g(r) = -\frac{GM_M}{|r|}, |r| > R_M. \tag{1.1} \]

Here \( G \) is the gravitational constant, \( M_M \) is the total mass of Mars, \( R_M \) is its mean radius, and \( r \) is the position of an arbitrary point in space relative the planet’s center of gravity. The associated gravitational force on an atmospheric particle of mass \( m \) at \( r \) is

\[ F_g(r) = -m\nabla V_g(r) = -m\frac{GM_M}{r^2}\hat{r}. \tag{1.2} \]

Escape requires non-zero kinetic energy after a particle has cleared the potential well which, in theory, occurs at an infinite distance. We may consider the work exerted on a particle starting near the planet at a position \( r_0 \), with an initial velocity \( v_0 \), following an arbitrary path \( C \) with infinitesimal steps \( ds \) from \( r_0 \) to an arbitrary point at infinite distance. The domain of valid escaping paths can be defined as

\[ \frac{m|v_0|^2}{2} + \int_C \left( F_{\text{other}}(r) - m\frac{GM_M}{r^2}\hat{r} \right) ds > 0, \tag{1.3} \]

where \( F_{\text{other}}(r) \) includes forces other than gravity, e.g. electromagnetic fields, which can accelerate charged particles. The path \( C \) also cannot intersect the collisional atmosphere. Solutions to this expression will be highly complex as the global flows of particles will change the vector field \( F_{\text{other}}(r) \) and thus the escaping domain in \( [r_0,v_0] \)-space.

In a simplified spherically symmetric system with no force other than gravity, the expression can be simplified to

\[ \int_{r_0}^{\infty} -m\frac{GM_M}{r^2} dr + \frac{mv_0^2}{2} > 0, \tag{1.4} \]

from which we can find the (minimum) escape velocity, \( v_{\text{esc}} \), as

\[ v_0 > \sqrt{\frac{2GM_M}{r_0}} = v_{\text{esc}}. \tag{1.5} \]

On the Martian surface (\( r_0 = 3390 \) km) the escape velocity becomes \( v_{\text{esc}} = 5.03 \) km/s. However, the Martian atmosphere is highly collisional at low altitudes, i.e. if a particle was moving with \( v > v_{\text{esc}} \), the probability to lose the associated energy through collisions with other atmospheric particles before reaching space is extremely high. At higher altitudes the collision probability decreases as the column density between \( r_0 \) and open space decreases. We can define the non-collisional regime (exosphere) as the range of altitudes, \( h \), where the mean free path of a particle,

\[ l_{\text{mfp}}(h) = \frac{1}{\sqrt{2n(h)\bar{\sigma}}}, \tag{1.6} \]

is larger than the atmospheric scale height.
Chapter 1. Introduction

\[ H(h) = \frac{kT(h)}{\bar{m}g(h)}. \]  

(1.7)

Here \( n \) is the number density of all gaseous species, \( \bar{\sigma} \) is the average collisional cross-section, \( \bar{m} \) is the average molecular/atomic mass of an atmospheric particle and \( g \) is the gravitational acceleration.

The exact altitude of the Martian collisional regime boundary (exobase) varies with season and solar-zenith angle in the typical range 140-200 km [Jakosky et al., 2017]. Note that in reality the boundary between the collisional- and non-collisional regime is not a discrete layer like the concept of an exobase implies, but rather a transition region of the length of one scale height, i.e. \( \sim 15 \) km.

Escape velocity at a typical exobase altitude of 170 km becomes 4.9 km/s. Different particle species carry different masses and as such require different energies to reach escape velocity. For example, an oxygen atom requires an escape energy of 2.0 eV, while a hydrogen atom will escape at only 0.13 eV. A table with escape energies for common particle species in the upper atmosphere is shown in Table 1.2.

All effective escape mechanisms ultimately shift, broaden or increase the density of the velocity distribution so that a higher absolute number of particles have energies and trajectories in the escaping domain of the local velocity space.

### Jeans escape

In the collisional atmosphere, the velocity distribution is determined by the gas temperature, thus the temperature and density of individual species at the exobase determine the corresponding rates of thermal (Jeans) escape to space. Here the velocity probability density distribution can be described by the Maxwell-Boltzmann (Maxwellian) distribution,

\[ f(v, m, T) = \sqrt{\frac{m}{2\pi k_B T}} \frac{3}{4\pi v^2} e^{-\frac{mv^2}{2\bar{m}k_B T}}, \]  

(1.8)

where \( T \) is the temperature, and \( k_B \) is the Boltzmann constant. The escaping fraction of the population can be found by integrating the probability density distribution over half (only outflowing) of all velocities higher than the escape velocity

\[ f_{esc}(m, T) = \frac{1}{2} \int_{v_{esc}}^\infty f(v, m, T)dv. \]  

(1.9)

Example distributions for H, H_2, O and CO_2 are shown in Figure 1.7 for a relatively hot 300 K Martian exosphere. The distributions illustrate well how Jeans escape is important for hydrogen escape at Mars, but not for heavy species. However, molecules with light constituents such as H_2O may still deplete through photodissociation and subsequent H-loss.

<table>
<thead>
<tr>
<th>Species</th>
<th>Escape energy [eV]</th>
</tr>
</thead>
<tbody>
<tr>
<td>H</td>
<td>0.13</td>
</tr>
<tr>
<td>He</td>
<td>0.50</td>
</tr>
<tr>
<td>N</td>
<td>1.8</td>
</tr>
<tr>
<td>O</td>
<td>2.0</td>
</tr>
<tr>
<td>O_2</td>
<td>4.0</td>
</tr>
<tr>
<td>CO_2</td>
<td>5.5</td>
</tr>
<tr>
<td>e^-</td>
<td>( 6.8 \times 10^{-5} )</td>
</tr>
</tbody>
</table>

Table 1.2: Escape energies for various particle species based on the escape velocity 4.9 km/s at typical exobase altitude (170 km).
1.2 Mechanisms for atmospheric loss

The fraction of H and H$_2$ that remain bound still reach significant altitudes, forming the Martian hydrogen corona, which exhibits variations in densities up to an order of a magnitude [Barabash et al., 1991; Chaffin et al., 2014; Rahmati et al., 2017; Yamauchi et al., 2015], larger than expected from Jeans theory alone. One proposed explanation is seasonal weakening of the Martian cold trap, allowing low-altitude water vapor to reach high altitudes and photodissociate [Chaffin et al., 2017]. Given that present-day estimates of H escape rates coalesce around $\sim5 \times 10^{26}$ s$^{-1}$ with seasonal maxima up to $10^{27}$ s$^{-1}$, and a doubling of the deuterium-to-hydrogen (D/H) ratio since the Hesperian, the thermal escape of H has played a significant role in the evolution of the Martian water inventory since. The total amount of H escaped to space corresponds to a 50-100 m equivalent global water layer [Mahaffy et al., 2015a; Krasnopolsky, 2015].

**Hydrodynamic escape**

The large difference in thermal/Jeans escape rate between light and heavy species implies there is a net flux difference between the two populations. The heavier particles in the atmosphere impart a drag force on the lighter species, which implies transfer of momentum to the heavier species. The flux of “dragged” heavy atoms (O, C, N) can be expressed using the formula by Hunten et al. [1987] in the formulation by Lammer et al. [2011] as

$$ F_O = \frac{X_O}{X_H} F_H \left[ \frac{m_H + \frac{k_B T H}{b_g X_H}}{m_H + \frac{k_B T H}{b_g X_H}} - m_O \right], $$

(1.10)

where $F_O$ and $m_O$ are respectively the flux and mass of the heavy particle species, oxygen in this example. $X_O$ and $X_H$ are the mole mixing ratios, $b$ is the oxygen diffusion parameter in hydrogen and $g$ is the gravitational acceleration.
The relative scarcity of light species and moderate H escape rates in the present Martian environment implies that the hydrodynamic drag effect is currently small. However, in the first 100 Ma after formation, under the extreme ultraviolet (EUV) heating and dissociation of the upper atmosphere by the young Sun, hydrodynamic escape is expected to have removed nearly all of the primordial steam-rich atmosphere [Lammer et al., 2012].

**Photochemical escape**

The absorption of X-ray and EUV photons from the Sun dissociates and ionizes molecules and atoms in the upper atmosphere, creating a rich mixture of neutral and ionized species that undergo a complex series of chemical reactions [Benna et al., 2015; Krasnopolsky, 2002]. Some of these reactions are exothermic, giving off reaction products with high energies relative to the background 200-300 K thermal distribution. Particularly important for escape of heavy constituents is the dissociative recombination of O\(^+\) ions with electrons, where the branching ratios (probabilities) for the reaction products and associated energy release are

\[
O_2^+ + e^- \rightarrow \begin{cases} O(3P) + O(3P) + 6.99 \text{ eV} & (22\%) \\ O(3P) + O(1D) + 5.02 \text{ eV} & (42\%) \\ O(3P) + O(1S) + 2.80 \text{ eV} & (<1\%) \\ O(1D) + O(1D) + 3.06 \text{ eV} & (31\%) \\ O(1D) + O(1S) + 0.83 \text{ eV} & (5\%) \end{cases}
\]

for the O\(^2_2^+\) vibrational ground state, as found through heavy ion storage ring experiments by Kella et al. [1997]. The energy released is shared equally between the hot O reaction products, from which follows that 62% of reactions will produce oxygen atoms with energies above escape velocity. This simplistic model is complicated by energy spread from distributions of vibrational states, altitude of production and temperature-dependent reaction rates

\[
R_O = 2\alpha_{DR}(T)n_{O_2^+}n_e^-,
\]

where a rate coefficient \(\alpha_{DR}(T) = 2 \times 10^{-7} (300K/T_e)^{0.7} \text{ cm}^3 \text{ s}^{-1} \) \(T < 1200 \text{ K}\) is suitable for the Martian ionospheric environment [Fox and Hać, 2009].

The hot O products create the Martian hot oxygen corona, which is tenuous, large in extent, and populated by both the bound and the escaping populations, which are closely separated at a low energy range. This combination of factors has so far hindered direct measurements of the photochemical escape. The last decade of models of the photochemical escape rate give rates \((0.9 - 6.0) \times 10^{25} \text{ s}^{-1}\) for solar minimum and \((2.1 - 19) \times 10^{25} \text{ s}^{-1}\) for solar maximum [Chaufray et al., 2007; Fox and Hać, 2014; Lee et al., 2015; Valeille et al., 2009].

Recently, with the arrival and accumulation of data by the Mars Atmosphere and Volatiles EvolutionN (MAVEN) orbiter from the U.S. National Aeronautics and Space Administration (NASA), a few studies have derived semi-empirical estimates for the photochemical escape rates. Rahmati et al. [2017] used measurements of solar wind accelerated/pick-up O\(^+\) (see Chapter 2), created from photoionized
coronal atoms, to derive the phase space distribution of the hot O corona. The corresponding escape rates were reported as within a factor 2 of $7 \times 10^{25} \text{s}^{-1}$.

Lillis et al. [2017] recently used MAVEN in-situ measurements of electron temperatures and O$_+^2$ densities to derive hot O production rates and inform an escape probability Monte-Carlo transport model, deriving escape rates $(1.2 - 5.5) \times 10^{25} \text{s}^{-1}$ for the time period Feb. 2015 to July 2016. Combined with Solar EUV measurements, and a model of solar EUV evolution, they also derive an overall escape rate $4.3 \times 10^{25} \text{s}^{-1}$ for the present-day Mars, as well as a strong escape rate dependence on CO$_2$ photoionization frequency with an exponent $2.6 \pm 0.6$. Extrapolated over time, this dependence would account for ca. 0.5 bar of atmosphere lost over the last 3.5 Ga (since the early-mid Hesperian). However, as noted by Cravens et al. [2017], hot O escape rates are sensitive to backscatter O–CO$_2$ cross-sections, for which current estimates are ill-constrained.

**Sputtered escape**

Sputtering is the collisional transfer of momentum from an energetic particle to a cold, bound, impacted particle. In the case of Mars, the energetic particles that enter the atmosphere are primarily charge-exchanged solar wind protons and solar wind accelerated "picked-up" O$^+$ ions. Solar Energetic Particles (SEPs) also enter the atmosphere, though the low mass of the electrons provides too small momentum transfer to initiate escape by sputtering [Leblanc et al., 2002].

The ~1 keV protons that make up the bulk of the solar wind are normally deflected around the planet by electric and magnetic forces. In the case mentioned above, a small fraction of the protons are neutralized by charge-exchange with coronal atoms and are thus able to enter the atmosphere. In the latter case, the O$^+$ ions are created from photoionized coronal oxygen atoms and accelerated up to several 10’s of keV by the motional/convective electric field of the solar wind. The forces that deflect the solar wind can not deflect particles with such high energies, thus they penetrate to the atmosphere. Solar wind interaction and acceleration processes are covered in further detail in the next subsection, as well as Chapter 2.

Sputtering escape rates are difficult to measure directly and thus current estimates rely on extensive use of models. Simulations by Wang et al. [2015] yield $6 \times 10^{23} \text{s}^{-1}$ for typical present-day EUV levels and $10^{26}$–$10^{27} \text{s}^{-1}$ for the early solar system. However, it is currently unclear how these estimates conform with observations and the strong dependence of pickup O$^+$ reflection fractions on $p_{\text{dyn}}$ recently reported by Masunaga et al. [2017].

**Ion escape**

In contrast to neutrals, ionized particles can be accelerated by electromagnetic forces generated by charge-separation in the planet’s ionized upper atmosphere, i.e. ionosphere, and through its interaction with the solar wind, providing a potential source of power to accelerate the atmospheric ions above escape velocity. The particles gain energy through acceleration by the local electric field, $\mathbf{E}$, which arises due to currents, $\mathbf{J}$, magnetic fields, $\mathbf{B}$ and the electron pressure gradient, $\nabla p_e$,

$$\mathbf{E} = -\mathbf{v} \times \mathbf{B} + \frac{1}{n_e e} \mathbf{J} \times \mathbf{B} - \frac{1}{n_e e} \nabla p_e,$$

(1.13)
Chapter 1. Introduction

The interaction with the solar wind results in generation of electric fields described by equation 1.13. However, the underlying processes are difficult to model accurately and will be presented in further detail in Chapter 2. Nevertheless, many studies have reported escape rates based on measurements.

The first direct measurements of atmospheric heavy ion fluxes were made by the Registratör Intensivnosti Elektronov Protonov (RIEP) cylindrical ion spectrometers on the Soviet Mars 2, Mars 3 and Mars 5 missions in 1971 (Mars 2, 3) and 1974 (Mars 5). The first estimate of the Martian ion escape rate was produced from the RIEP measurements and reported by Bogdanov et al. [1975] as $\sim 10^{25}$. [Vaisberg et al., 1977] later took $10^{25}$ s$^{-1}$ as an upper estimate and a review by Vaisberg et al. [1986] gives a typical escape rate $10^{24}$ s$^{-1}$.

Phobos-2, another Soviet orbiter, arrived at Mars in February 1989. The onboard Automatic Space Plasma Experiment with a Rotating Analyzer (ASPERA) instrument comprised a top-hat electron spectrometer and an ion mass spectrometer, which performed nearly two months of measurements in Martian orbit, including four highly elliptic orbits from which Lundin et al. [1990] estimated an ion escape rate $3 \times 10^{25}$ s$^{-1}$.

The European Space Agency (ESA) inserted Mars Express (MEX) in orbit of Mars in December 2003. Barabash et al. [2007] calculated the first ion escape rate from MEX measurements as $4 \times 10^{23}$ s$^{-1}$, using data from the Analyzer of Space Plasmas and Energetic Atoms (ASPERA-3) particles package. The low rate reported by Barabash et al. [2007] is due to the relatively high 30 eV lower energy limit of the Ion Mass Analyzer (IMA) energy table used early in the mission, and a cold ion escape population later discovered to be dominant close to the planet. Escape rates reported after upload of new low-energy IMA energy tables in 2007 (>10 eV) and 2009 (>1 eV), were in the range $2.0 \times 10^{24}$ s$^{-1}$ to $3.3 \times 10^{24}$ s$^{-1}$, as measured close to solar minimum [Lundin et al., 2008b; Nilsson et al., 2011].

MAVEN is the most recent orbiter to provide ion flux measurements at Mars. Ion escape rates measured by the onboard SupraThermal And Thermal Ion Composition (STATIC) instrument have been reported as $3 \times 10^{24}$ s$^{-1}$ (>25 eV) [Brain et al., 2015] and $(2–3) \times 10^{24}$ s$^{-1}$ (>6 eV, only O$^+$) [Dong et al., 2017].

Across different missions and time, the reported ion escape rates have thus remained remarkably stable in the range $\sim 10^{24}$–$10^{25}$ s$^{-1}$, which are unable to account for atmospheric losses over a few 10’s of mbar. However, studies of other G–type stars of varying ages show that the young (3.9 Ga) Sun was significantly more active, producing on average an order of magnitude higher X-ray and EUV intensities [Ribas et al., 2005], and mass-loss rates close to 2 orders of magnitude higher [Wood, 2006]. The higher solar wind flux provides more power available to accelerate ions past escape velocity, and increased solar radiation enhances the production rate of ions through photoionization. However, the interaction between the solar wind and the Martian ionosphere is complicated and it is currently not clear which factors are limiting the ion escape rate. Finding the escape trends and associated cumulative contribution to the atmospheric pressure lost since the collapse of the Martian hydrosphere in the Hesperian has been the ultimate goal of the work behind this thesis.
2. Solar wind interaction with Mars

2.1 The Sun, solar wind and interplanetary magnetic field

2.1.1 Properties of the solar wind

The solar wind is a stream of highly ionized plasma driven by the expansion of the solar corona throughout interplanetary space. The coronal plasma at the source region near the Sun is highly magnetized due to the strong magnetic fields originating from the Solar interior, and the currents that connect the Sun to the rest of the solar system. As noted by Alfvén [1957] and Parker [1958], the magnetic fields are thus “frozen in” the expanding plasma, forming the Interplanetary Magnetic Field (IMF). The rotation of the Sun and expansion of the solar wind together gives the IMF a spiral-like shape, i.e. the Parker spiral, as illustrated in Figure 2.1 together with the Martian heliocentric orbit.

In the frozen-in condition, currents and polarization gradients are small, thus the electric field, as given by eq. 1.13, is dominated by the motional component

$$E_{\text{mot}} = -v_{\text{sw}} \times B_{\text{IMF}}.$$  \hspace{1cm} (2.1)

The Lorentz force equation shows that this is a force-free equilibrium,

$$\mathbf{F} = q(\mathbf{E} + \mathbf{v} \times \mathbf{B}),$$  \hspace{1cm} (2.2)

thus individual ions and electrons drift without gyration from a stationary observer’s perspective. In the solar wind comoving frame $v'_{\text{sw}} = 0$, thus $E'_{\text{mot}} = 0$. The expanding coronal plasma thins with increasing heliocentric distance as $1/r^2$ and is primarily composed of H$^+$ ($\sim 97\%$) and He$^{2+}$ ($\sim 3\%$), with contributions from heavier ions at high charge states (e.g. O$^{6+}$–O$^{8+}$, C$^{4+}$–C$^{6+}$, Mg$^{8+}$–Mg$^{10+}$, Fe$^{8+}$–Fe$^{16+}$) generally below 0.1\% [Lepri et al., 2013]. Typical upstream solar wind and IMF parameter values at Mars are density $n_{\text{sw}} = 2.5 \text{ cm}^{-3}$, velocity $v_{\text{sw}} = 400 \text{ km/s}$, temperature $T_{\text{sw}} = 4 \times 10^4 \text{ K}$ and $B_{\text{IMF}} = 2 \text{ nT}$, with a spiral (cone) angle of $\sim 56^\circ$ providing a motional electric field $E_{\text{mot}} \approx 0.7 \text{ mV/m}$. The low proton thermal speed to bulk speed gives the solar wind a typical angular width $\tan^{-1}(v_{\text{thermal}}/v_{\text{bulk}}) \approx 2^\circ$, i.e. a relatively cold, fast beam.
Chapter 2. Solar wind interaction with Mars

Figure 2.1: Illustration of the Martian orbit and orientation in the rotating heliosphere, sizes of the bodies are not to scale. Solar longitude zero, \( L_S = 0^\circ \), is defined by the Martian vernal equinox, placing the orbit aphelion at \( L_S = 71^\circ \), i.e. near summer solstice in the northern hemisphere. Solar wind and EUV radiation are emitted radially from the Sun and are thus regulated by up to 46% from the eccentric Martian orbit alone. The IMF is frozen into the solar wind flow near the Sun and thus takes the shape of a spiral due to the flow of the solar wind and ~27 day rotation period of its source regions.

In reality, at any given time each upstream parameter at Mars falls on a distribution shaped by the 11-year solar activity cycle and the eccentric Martian orbit, modulating the heliocentric distance. Fast solar wind streams emanating from coronal holes, i.e. open solar field lines, can overtake slower streams and create compression regions at the stream interface known as Corotating Interaction Regions (CIRs), as well as rarefaction regions in the wake. Similarly, fast Coronal Mass Ejections (CMEs), strongly magnetized clouds of accelerated solar plasma, can pile up solar wind and leave rarefied wakes. Both types of events are products of the complicated magnetic topology created near the solar surface as the Sun flips polarity and are typically more developed at the Martian orbit compared to at the planets closer to the Sun, e.g. Earth. Mars Express has provided the longest period of in situ solar wind measurements at Mars, albeit with an instrument not designed to detect protons over the full energy range of the solar wind (see Chapter 3). Developing accurate solar wind moments from Mars Express measurements has been a significant part of the project outlined in this thesis and the resulting \([n_{sw}, v_{sw}]\) distribution is shown in Figure 2.2.

2.1.2 Solar radiation
The solar cycle also modulates the intensities of solar X-ray (1–10 nm) and EUV (10-124 nm) radiations that heat and ionize the upper Martian atmosphere. The 1–124 nm range comprises a forest of emission lines and various continua produced primarily in the corona and chromosphere-corona transition region. The Solar Extreme Ultraviolet Explorer instrument on the Thermosphere Ionosphere
2.1 The Sun, solar wind and interplanetary magnetic field

Figure 2.2: Distribution of solar wind density and velocity at Mars based on Mars Express measurements over nearly a full solar cycle (2007-2017). Here each measurement is based on data averaged over the ∼40 min of measurements taken before/after MEX enters/exits the Martian bow shock on most orbits.

Mesosphere Energetics and Dynamics spacecraft (TIMED/SEE) has provided X-ray and EUV measurements at Earth since 2002, i.e. for the full duration of the MEX mission [Wood et al., 2005]. Figure 2.3 shows example SEE spectra from three days representative of solar minimum, early maximum and peak maximum in solar cycle #24. Prominent spectral peaks corresponding to atomic emission lines can be seen at 17.1 nm (Fe IX), 30.4 nm (He II), 58.4 nm (He I), 97.7 nm (C III), 33.5 nm (Fe XVI) and 121.6 nm (Ly-α). Below 5 nm X-rays are produced by bremsstrahlung from free electrons colliding with ions, recombination emissions contribute continua at 45–50 nm and 70–90 nm, corresponding to the sum of the initial electron kinetic energy and the ionization/recombination energy.

Figure 2.3 shows that solar activity strongly affects short wavelength emissions (λ ≲ 38 nm), with small effects on longer wavelengths, which are important for ion production rates in the Martian atmosphere as the photon flux depends on irradiance and photon energy as

\[ F_\gamma(\lambda) = \frac{I(\lambda)}{E_\gamma(\lambda)} = \frac{I(\lambda)\lambda}{hc}. \]  

(2.3)

However, all wavelengths are affected equally by the varying distance to the Sun, imposed by the eccentric Martian orbit. The TIMED/SEE spectra can be propagated to Mars’ concurrent position in its orbit to reveal variations in total 1-118 nm flux with a dynamic range of ∼2 over the time period 2004–2017, as shown in Figure 2.4.
Figure 2.3: X-ray and EUV intensity spectra as measured by TIMED/SEE at 1 AU with the corresponding emission regions on the Sun as observed by the Solar Dynamics Observatory Atmospheric Imaging Assembly (SDO/AIA) on 2015-02-05. The three spectra show the difference between solar minimum (Jan. 2009), early maximum (Dec. 2011) and peak maximum (Feb. 2015, corresponding to the Sun images).

2.2 Mars upper atmosphere and ionosphere

The low-altitude Martian atmosphere is dominated by CO$_2$, at high altitudes the CO$_2$ and minor background species are photodissociated by a wide range of solar X-ray and EUV photons, producing a diverse altitude-dependent mixture composed of the dissociated products and subsequent chemical reaction products. The volume photoproduction rate of a particular species Y, $R_Y$, is the sum of photoreactions of parent species, X, with resultant child species Y, i.e.

$$R_Y = \sum_X \int_0^\infty F_Y(\lambda) \sigma_{X\rightarrow Y}(\lambda)n_Xd\lambda.$$  \hspace{1cm} (2.4)

Here, $F_Y$ is the photon flux, $\sigma_{X\rightarrow Y}$ is the photo cross-section of parent species X with child species Y, and $n_X$ is the number density of parent species X. Photo cross-sections for the most common C– and O–bearing parent species in the Martian upper atmosphere are shown in Figure 2.5.

CO$_2^+$ photoions are produced in a variety of quantum states which absorb specific energy quanta, e.g. 13.77 eV ($X^2\Pi_u$), 17.32 eV ($A^2\Pi_u$), and 18.10 eV ($B^2\Sigma_u^+$). The strong absorption of the solar 30.4 nm He II line, in combination with...
2.2 Mars upper atmosphere and ionosphere

Figure 2.4: Daily average X-ray and EUV photon fluxes as measured at Earth by TIMED-SEE and propagated to Mars for the duration of the MEX mission. Upper panel shows flux per spectral 1 nm bin, lower panel shows photon fluxes integrated over separate spectral intervals. Several quasi-periodic variations can be seen driven by the 11-years solar activity cycle, the 1.88-year period of the eccentric Martian orbit, and the ∼27-day Carrington rotation period of the Sun.

a high photoionization cross-section, results in distinct 20–27 eV photoelectrons observed in the dayside ionosphere and even in the far magnetotail [Mantas et al., 1979; Frahm et al., 2006], indicating that field lines in the tail are often connected to the dayside ionosphere. Overall, the absorption of photoionizing radiation and corresponding photon attenuation results in a double-peaked plasma density profile corresponding to EUV and X-ray absorption, the M2 and M1 layers, respectively (e.g. Withers et al. [2012]). The main peak altitude is nominally located at altitudes 120–140 km, though higher for solar-zenith angles (SZA) larger than 70°–80° [Sánchez–Cano et al., 2016]. The upper extent of the thick thermal ionosphere is marked by a relatively sharp ionopause near 400 km altitude, however, the thinner top-side ionosphere extends to significantly higher altitudes as apparent from the extent of photoelectrons [Han et al., 2014]. The exact location of the upper boundary for the top-side ionosphere is strongly dependent on solar wind and EUV conditions [Ramstad et al., 2017a].

Besides direct photodissociation and photoionization, a number of subsequent
Chapter 2. Solar wind interaction with Mars

Figure 2.5: Photodissociation and photoionization cross-sections [Huebner et al., 2015] for major carbon- and oxygen-bearing species in the Martian upper atmosphere (CO₂, O₂, CO, and O, in respective panels). Here only the ground states of the parent species are included and cross-sections are summed over all states of the child species.

Chemical reactions alter the composition and work to dissociate CO₂ to lighter components, a few of the most important reactions in the Martian ionosphere are

\[ \text{O}^+ + \text{CO}_2 \rightarrow \text{O}_2^+ + \text{CO} \]  
\[ \text{C}^+ + \text{CO}_2 \rightarrow \text{CO}^+ + \text{CO} \]  
\[ \text{CO}^+ + \text{CO}_2 \rightarrow \text{CO}_2^+ + \text{CO} \]  
\[ \text{CO}_2^+ + \text{O} \rightarrow \text{O}_2^+ + \text{CO} \]

with subsequent breakdown of the child species e.g. through dissociative recombination (see eq. 1.11). An extensive overview of the Martian upper atmosphere reactions and rate coefficients is available e.g. from Krasnopolsky [2002].

The Neutral Gas and Ion Mass Spectrometer (NGIMS) instrument on MAVEN has, since arrival in late 2014, provided the highest mass-resolution in situ measurements of the neutral upper atmosphere. Figure 2.6 shows neutral profiles reported by Mahaffy et al. [2015b]. The CO₂-dominated domain can clearly be seen to extend up to ca. 250 km altitude, above which O and O₂ dominate.

The photodissociation and chemical reaction chains result in ionospheric composition profiles significantly different from the neutral atmosphere, NGIMS ion density measurements reported by Benna et al. [2015] are shown in Figure 2.7. O₂⁺ dominates at low altitudes and can be seen to form a ~50/50 mix with O⁺ at altitudes ≥300 km with relative concentrations \( n_{\text{CO}_2^+} / n_{\text{O}^+} \lesssim 10\% \). The typical composition of the escaping heavy ions roughly reflects this high-altitude composition [Carlsson et al., 2006; Nilsson et al., 2011].
2.2 Mars upper atmosphere and ionosphere

Figure 2.6: Density profiles of neutral species in the Martian upper atmosphere as measured by the NGIMS instrument on MAVEN near local noon (11:50 AM LST). Adapted from Mahaffy et al. [2015b] (CC-BY-NC-ND).

Figure 2.7: Density profiles of ionospheric species at Mars as measured by the NGIMS instrument on MAVEN at ca. 60° solar-zenith angle. Adapted from Benna et al. [2015], with permission.
2.3 The Martian induced magnetosphere

The Martian ionosphere forms a conductive obstacle to the solar wind. We can describe the interaction at the boundary through the magnetohydrodynamic (MHD) momentum equation

$$\rho \left( \frac{\partial}{\partial t} + \mathbf{v} \cdot \nabla \right) \mathbf{v} = -\nabla p + \mathbf{J} \times \mathbf{B}, \quad (2.6)$$

where \( \rho \) is the plasma density, \( \nabla p \) is the thermal pressure gradient and \( \mathbf{J} \times \mathbf{B} \) is the Hall force contribution arising from the current density \( \mathbf{J} \) and magnetic field \( \mathbf{B} \). Neglecting the displacement current term of the Maxwell-Ampère equation, assuming a low-frequency interaction, we may take \( \mathbf{J} = \frac{(\nabla \times \mathbf{B})}{\mu_0} \). Under steady-state conditions, i.e. \( \partial / \partial t = 0 \), the momentum equation thus becomes

$$\rho (\mathbf{v} \cdot \nabla) \mathbf{v} + \nabla p - \frac{1}{\mu_0} (\nabla \times \mathbf{B}) \times \mathbf{B} = 0. \quad (2.7)$$

The induced magnetic force term can be broken up in two quantities,

$$\frac{1}{\mu_0} (\nabla \times \mathbf{B}) \times \mathbf{B} = -\nabla \left( \frac{B^2}{2\mu_0} \right) + \frac{1}{\mu_0} (\mathbf{B} \cdot \nabla) \mathbf{B}, \quad (2.8)$$

i.e. the magnetic pressure gradient and the magnetic tension force. Assuming \( \mathbf{B} \) only changes across the direction of \( \mathbf{B} \), so that \( (\mathbf{B} \cdot \nabla) \mathbf{B} = 0 \), we are left with the gradient of the magnetic pressure. Thus eq. 2.9 becomes

$$\rho (\mathbf{v} \cdot \nabla) \mathbf{v} + \nabla p + \nabla \left( \frac{B^2}{2\mu_0} \right) = 0. \quad (2.9)$$

In the case of the solar wind we may take the density \( \rho = \rho_{sw} \) and \( \mathbf{v} = -v_{sw} \cos(\theta) \hat{x} \), where \( \theta \) is the angle between the solar wind flow and the boundary normal vector \( \hat{x} \). At the interface between the solar wind and ionosphere we are left with a 1D pressure balance equation, assuming incompressible flow,

$$\frac{\partial}{\partial x} \left( \frac{1}{2} \rho_{sw} v_{sw}^2 \cos^2 \theta + p_{th} + \frac{B^2}{2\mu_0} \right) = 0, \quad (2.10)$$

where \( \rho_{sw} v_{sw}^2 / 2 \) is the solar wind dynamic pressure, \( p_{th} \) is the thermal pressure and \( B^2 / 2\mu_0 \) is the magnetic pressure. Neglecting the relatively small thermal pressure term, under equilibrium the interaction can be simplified as

$$\frac{1}{2} \rho_{sw} v_{sw}^2 \cos^2 \theta + \frac{B^2}{2\mu_0} = \text{constant}. \quad (2.11)$$

In other words, the induced magnetic pressure balances the solar wind dynamic pressure along a streamline, halting and deflecting the solar wind at a location often named the Induced Magnetosphere Boundary (IMB). The induced magnetic pressure thus depends on the solar wind dynamic pressure and the local flaring angle \( \theta \).
The induced magnetic fields created by the currents in the ionosphere act on the solar wind protons, which gyrate in the magnetic barrier region and are turned back into the incoming flow. Interaction of the two beams result in instabilities and waves which increase the plasma temperature and thus reduce the Mach number below 1. Superposition of the upstream-propagating magnetosonic waves results in a shock. The plasma passing through the bow shock is heated and decelerated. Here the field lines move slower than the ends in the undisturbed solar wind. The magnetic fields stretch and drape around the obstacle. Since the solar wind is supersonic, an obstacle creates a wake, a solar wind void. The draping fields maintain the pressure balance at the boundaries of the wake creating a magnetic tail. Note that this configuration forms for any highly conductive obstacle immersed in a supersonic plasma flow with a frozen-in magnetic field. The motional electric field creates a potential across the tail which drives a cross tail current closing along the wake boundaries. The resulting $J \times B$-force extracts and accelerates ions from the night side ionosphere in a plasma sheet [Yeroshenko et al., 1990; Dubinin et al., 2017].

The Martian dayside ionosphere is often magnetized and connected to the tail as evident from the presence of non-intrinsic magnetic flux tubes in the dayside ionosphere and presence of photoelectrons in far downtail regions [Dubinin et al., 2014; Frahm et al., 2006]. Thus, electron temperatures and heating rates in the connected dayside ionosphere are important as ionospheric outflow may be driven by a polarization electric field,

$$E_\parallel = -\frac{k_B T_e}{e} \frac{1}{n_e} \frac{\partial n_e}{\partial s}, \quad (2.12)$$

where $s$ is an arbitrary position on the connected field line, and $n_e, T_e$ are respectively the electron density and temperature, here neglecting the contribution from the photoelectrons themselves. An upper limit on the Martian ambipolar electric field has been derived by Collinson et al. [2015] through studies of changes in the well-known characteristic energies of the photoelectrons, indicating that the potential difference is less than 4.5 V in the magnetotail and $\lesssim 2$ V in the top-side ionosphere. Although these estimates are upper limits, a potential difference on the order of only a few volts would be capable of driving an ionospheric polar outflow along the open field lines (compare with escape energies in Table 1.2), producing the cold ion outflow often observed in the tail [Dubinin et al., 2011]. A schematic illustration of the Martian induced magnetosphere is shown in Figure 2.8.

It should be noted that the above MHD description for the Mars solar wind interaction can only serve as an approximation as all distributions are not Maxwellian and the gyroradii of heavy atmospheric ions are in large areas on scales similar to the size of the obstacle. In particular, penetration of the solar wind electric field (eq. 2.1, appears as a convective field in the magnetosheath) in scavenging regions, in addition to photoionization of the Martian exosphere, leads to the presence of heavy ions in the magnetosheath that are picked up by the convective electric field, forming an energetic heavy ion "plume" [Dong et al., 2015; Liemohn et al., 2014].
Figure 2.8: Schematic illustration of the Martian induced magnetosphere and ionosphere interaction with the solar wind. The frozen-in IMF induces currents in the upper ionosphere, repelling the solar wind flow at the IMB, forming a bow shock at the transition to subsonic speeds. The stretching and draping of the IMF forms a magnetotail, with two lobes of opposite polarity separated by a current sheet where ions are accelerated by the $\mathbf{J} \times \mathbf{B}$-force. The Martian top-side ionosphere is often magnetized as shown here, with field-lines connected to the tail, through which the ionosphere can outflow in the presence of a polarization electric field.

2.4 Crustal fields

The 1965 flyby of Mariner IV provided the first indications that Mars lacked an Earth-like intrinsic global magnetic dipole [Smith et al., 1965], however for the next three decades a significant upper limit remained. Magnetometer data from the initial aerobraking and phasing orbits of MGS showed no evidence for a global dipole but rather widespread strong crustal magnetization, particularly while the spacecraft was passing over the oldest terrains [Acuña et al., 1999].

A geographical map of the Martian crustal fields is shown in Figure 2.9, based on the empirical spherical harmonics model by Morschhauser et al. [2014], in turn created from MGS Magnetometer/Electron Reflectometer (MAG/ER) data. The crustal fields are mainly present in the ancient Noachian regions, appearing with banded polarities, particularly in Terra Cimmeria and Terra Sirenum. On Earth, such banded polarity structures spread out from sea-mounts on the ocean floor where new crust solidifies and is magnetized by the Earth’s global magnetic field as the new material cools below the Curie temperature. The banded Martian crustal fields are thus indicative of similar tectonics, and the presence of an ancient global dipole, on early Mars.
The precise strength of the ancient dipole can not be well constrained from the crustal fields alone as the crustal magnetization depends also on the remanence, concentration and distribution of the ferromagnetic crustal materials. However, the estimated near-surface field over the strongest magnetized region (178°E, 53°S) has been estimated as 15,800–19,900 nT, depending on the assumed source depth [Brain et al., 2003]. For comparison, the strongest crustal fields on the Earth’s surface are ~500 nT [Lesur et al., 2016] and the global dynamo-generated dipole strength 25,000–65,000 nT, depending on geomagnetic latitude. The largest and youngest Martian impact basins have been nearly completely demagnetized by impact shock and heating above the Curie temperature (Argyre, Hellas, Isidis, Utopia) while older impact basins have remagnetized (e.g. Amazonis, Amenthes, Ares, Deadalia, Sirenum). Comparing ages and magnetization of the impact basins suggests that the ancient global dipole collapsed abruptly (within 20 Ma) in the early Noachian, 4.12 Ga ago, or ca. 450 Ma after formation [Lillis et al., 2008]. Thus the planet has existed in an overall non-magnetized state for most of the Noachian and all of the time since, with the exception of the crustal fields.

The crustal fields alter the interaction with the solar wind by introducing intrinsic magnetic pressure, and when oriented on the dayside can locally offset the balance between the solar wind dynamic pressure and ionospheric induced magnetic pressure (eq. 2.11). The strongest fields can locally raise the solar wind stand-off distance and act as additional obstacles to the shocked solar wind flow in the magnetosheath [Crider, 2004; Edberg et al., 2009; Wang et al., 2014]. The crustal fields, in effect, act as miniature magnetospheres with many features in
common with their global dynamo-generated counterparts, including trapped and dispersed plasma [Brain et al., 2007; Harada et al., 2016], cusp-like open field line regions [Brain et al., 2007], auroral acceleration features [Lundin et al., 2006], flux tubes and likely reconnection [Brain et al., 2010].

The orientation of the crustal fields determines their influence on the the solar wind interaction. In areas where the crustal field strength is large enough to dominate the draped fields in the sheath (effectively acting as obstacles), the interaction strongly heats the ionosphere, though only for crustal field orientations SZA > 60° [Andrews et al., 2015]. For SZA < 60° there is little detectable heating, suggesting the crustal fields mainly offset the dynamic pressure of the solar wind. A similar relation to SZA is reflected in the cold ion outflow [Ramstad et al., 2016], discussed further in Chapter 4. An illustration of the possible crustal field–solar wind interaction for varying SZA is shown in Figure 2.10.

![Figure 2.10](image)

Figure 2.10: Cartoon illustration of the possible role of the crustal fields (green curves) in the interaction with the solar wind (stream lines shown) for varying orientations. a) SZA < 60°: The strongest fields locally offset the induced magnetosphere–solar wind pressure balance, increasing the stand-off distance, locally screening the ionosphere. b) SZA > 60°: The crustal fields extend further into the sheath due to lower local solar wind dynamic pressure at the boundary (eq. 2.11) and pose obstacles to the solar wind flow in the sheath (indicated by red-shaded area), displacing the IMB from its nominal location (dotted curve). The associated energy transfer heats the topside ionosphere [Andrews et al., 2015], driving tailward plasma outflow through flux ropes that may undergo reconnection [Brain et al., 2010].
3. Mars Express/ASPERA-3

3.1 Mars Express

The European Space Agency (ESA) approved the Mars Express (MEX) mission concept in November 1998, shortly after the failure of the Russian Mars-96 mission which carried several contributions from European institutions. The rapid development, from which the mission derives its name, resulted in a launch in June 2003 from Baikonur, Kazakhstan, and a successful Mars orbit insertion (MOI) in December the same year, just over 5 years after mission approval.

Mars Express originally consisted of the orbiter and a lander, Beagle 2, which failed to send communications after landing due to deployment failure of a solar panel, obstructing the communications antenna. The orbiter remains in good health as of 2017 and carries a diverse suite of scientific instrument packages [Wilson and Chicarro, 2004].

ASPERA-3 Analyzer of Space Plasmas and Energetic Atoms
HRSC High Resolution Stereo Camera
MaRS Mars Radio Science Experiment
MARSIS Mars Advanced Radar for Subsurface and Ionosphere Sounding
OMEGA Observatoire pour la Minéralogie, l’Eau, les Glaces et l’Activité
PFS Planetary Fourier Spectrometer
SPICAM Spectroscopy for the Investigation of the Characteristics of the Atmosphere of Mars

A schematic illustration of Mars Express is shown in Figure 3.1, including locations of the scientific instrumentation. The spacecraft main body is rectangular-like in shape with a slanted side, measuring 1.7/1.5 m × 1.7 m × 1.4 m with the 1.6 m diameter High-Gain Antenna (HGA) mounted on the slanted side. Dark Multi-Layer Insulation (MLI) in combination with a cluster of electrical heaters maintain internal operational temperatures throughout the Martian orbit. The solar
arrays collect sunlight over an effective area of 11.42 m² and pivot to maintain a high aspect angle to the Sun. A photograph of the spacecraft body under Thermal Balance/Thermal Vacuum (TBTV) testing is shown in Figure 3.2, highlighting the locations of the ASPERA-3 instrument units.

Figure 3.1: Mars Express spacecraft dimensions, reference axes and scientific instrumentation. The studies undertaken in this thesis mainly utilize the ASPERA-3 particles package, its components; the Main Unit (MU) and Ion Mass Analyzer (IMA), are highlighted in red and described in further detail in section 3.2. Note that the solar arrays pivot on their mounted axis, the top-side mounted Beagle 2 lander separated from Mars Express before MOI and the two 20 m MARSIS dipole antennas are not shown.

Mars Express studies the planet from a highly inclined (86.9°) elliptical nominal orbit with a 7.0 h period and 330 km (0.1 RM) peri-/apoareion altitudes, respectively. The orbit precesses due to relatively large asymmetries in the Martian gravity field, over time allowing the spacecraft to cover the near-Mars space over a wide orbit distribution, shown in Figure 3.3.

The spacecraft is 3-axis stabilized by means of reaction wheels which are regularly offloaded by the attitude thrusters. Close to the planet (±40 min from periareion) attitude varies from orbit to orbit depending on which instrument has pointing priority. Otherwise the HGA is kept pointed at Earth to maximize time for data uplink.
3.1 Mars Express

Figure 3.2: Mars Express undergoing Thermal Balance/Thermal Vacuum (TBTV) testing in 2003. The ASPERA-3 Main Unit (MU) and Ion Mass Analyzer (IMA) instruments are clearly visible. Note that the solar arrays were not mounted during the TBTV tests.

Figure 3.3: Mars Express orbit distribution (Jan. 2004–Dec. 2017) in cylindrical MSO coordinates. The day/night-sides of the planet as well as nominal Bow Shock, Induced Magnetosphere Boundary and Ionosphere Boundary models for nominal upstream conditions [Ramstad et al., 2017a] are shown for reference. Black areas indicate no coverage.
3.2 ASPERA-3

The Analyzer of Space Plasmas and Energetic Atoms (ASPERA-3) particles package was designed and built by a consortium led by the Swedish Institute of Space Physics/Institutet för Rymdfysik (IRF) in Kiruna, Sweden [Barabash et al., 2006].

Two units, carrying a total of four instruments, comprise ASPERA-3. The Main Unit (MU) is mounted on the $+zs/c$ plate, with the sensor heads and Digital Processing Unit (DPU) placed on turnable scanner platform that allows pointing of the constituent instruments; the Electron Spectrometer (ELS), Neutral Particle Detector (NPD) and Neutral Particle Imager (NPI). The Ion Mass Analyzer (IMA) is a separate unit and instrument mounted directly on the $−zs/c$ plate, connected to the MU by cable. Figure 3.4 shows both units as assembled and their placements on MEX is illustrated in Figure 3.1. IMA and ELS are described in detail below, however NPD and NPI will not be covered further here as these instruments were not used in the work behind this thesis.

![Figure 3.4: ASPERA-3 Ion Mass Analyzer (IMA) and Main Unit (MU) flight models. The individual instruments on the MU are indicated; Neutral Particle Detector (NPD), Neutral Particle Imager (NPI), and Electron Spectrometer (ELS). Red protective covers and purging gas lines are mounted on the NPD apertures, as well as a red cover mounting ring around NPI.](image)

3.2.1 Ion Mass Analyzer (IMA)

IMA was designed to measure the local 3D ion differential flux and mass distributions as the spacecraft passes through the various Martian plasma environments. The ion-optics assembly consists of a scanning electrostatic deflection system which sweeps entrance elevation angles $±45^\circ$ for a $127^\circ$ top-hat electrostatic energy analyzer (ESA) with an energy resolution 7%. The nearly monoenergetic beam exiting the ESA is subsequently accelerated by a post-acceleration ion lens, increasing the beam energy by a set 300, 2433 or 4216 eV (denoted PAC-0/4/7, respectively) before it enters a toroidal magnetic field which separates the beam’s constituent ions species by gyroradius. The gyroradius is proportional to the velocity of each ion species and thus gives the equivalent mass-per-charge ($M/q$). The distribution is measured by a microchannel plate (MCP), segmented in 16 azimuthal sector
anodes and 32 radially separated mass-rings. A schematic cutaway/cross-section
drawing of IMA and its ion optics assembly is shown in Figure 3.5.

The energy table used by IMA has changed over the course of operations. By
design, the energy table is always 96 levels, which are typically logarithmically
separated and can be set within the ESA’s 1 eV – 36 keV range. Between launch
in 2003 and April 30, 2007 the nominal energy table covered the range 30 eV –
32 keV. A software patch enabled better coverage in the low energy range from
May 2007, providing a new table from 10 eV to 25 keV. The latest change to the
energy table was implemented in November 2009, IMA has since maintained an
energy range 1 eV – 15 keV. Note that all top-hat ESAs technically select ions by
energy-per-charge, $E/q$. However, in the Martian environment, multiple charged
ions are rare and so simply energy is often used unless ions with higher charge
numbers are studied specifically, e.g. the solar wind alpha particles ($\text{He}^{2+}$).

The ion optics design and high-voltage supply settings provide an instantaneous
azimuth-elevation field of view (FOV) $360^\circ \times 4.5^\circ$, expanded to $360^\circ \times 90^\circ$ through
stepping the deflection voltage in 16 steps. However, the energy-elevation voltage
tables after the May 2007 patch have the deflection system effectively offline for
energies <50 eV, giving the instrument a flat FOV with an elevation solely determined by the energy-dependent entrance acceptance angle of the ESA. Also, as the instrument is mounted directly on the spacecraft body, a significant fraction of the FOV is blocked by surfaces. The resulting coverage of the instantaneous differential flux distribution is complicated and has to be accounted for when analyzing measurements. Figure 3.6 shows IMA FOV ray-traced blockage, developed as part of this project in order to improve estimation of solar wind moments.

Figure 3.6: IMA field-of-view in the instrument’s azimuth-elevation reference frame. Blockage (black areas) is found by backwards ray-tracing particle entry paths with surfaces on a 3D-model of Mars Express, here with the pivoting solar arrays oriented in the $x_s/c$-$y_s/c$-plane, i.e. near the most disadvantageous orientation for IMA. The grid shows the bins of individual azimuthal sectors (numbered) as scanned in elevation by the instrument’s deflection system.

Ions of different masses are separated by $M/q$-dependent gyroradii over the 32 radially separated MCP anodes (mass-rings). The energies of ions selected by the ESA at an arbitrary energy level have a FWHM 7% of the center energy, creating a distribution of gyroradii also for ions of the same $M/q$. Thus each ion species is spread out over several mass-rings and distributions of different species can overlap. Both the center position and spread of an ion distribution on the MCP depends on the energy level and the PAC-setting used. The center mass position dependence on energy for a species is called its mass-line or mass-curve. The separation of ionospheric species, $O^+$, $O^+_2$ and $CO^+_2$, is not straightforward and requires special techniques, particularly at high energies due to small relative differences in $M/q$ and converging mass-lines. An example IMA Mass-Energy matrix is shown in Figure 3.7. Ion species in IMA data are separated by fitting multiple Gaussian functions centered on the mass-lines throughout the studies detailed in this thesis.

The counts detected by a particle spectrometer such as IMA are proportional (ideally) to the incident flux within a given time, $\Delta t$, over an effective aperture area, $A^*$, from a range of space angles, $\Delta \Omega$, over a range of energies, $\Delta E$, and with a certain efficiency, $\eta$. Given knowledge of all these parameters, the registered
counts, $c$, can be converted to fully differential flux as

$$j = \frac{c}{G_E \Delta E},$$

where $\Delta t = 120.9$ ms is the integration time per energy step and

$$G_E = \Delta \Omega A^* \eta \frac{\Delta E}{E},$$

is the effective energy-geometrical factor ($cm^2$ sr eV/eV), which accounts for the energy resolution of the ESA, $\frac{\Delta E}{E} = 0.07$. The value for $G_E$ is PAC- and energy-dependent, derived from ground calibrations at Centre d’Etude Spatiale des Rayonnements (CESR), in Toulouse, France, since renamed as the Institut de Recherche en Astrophysique et Planétologie (IRAP). The provided ground-calibrated values for $G_E$ for heavy ions and the PAC-4 setting are shown in Figure 3.8. $G_E$ depends on energy due to different trajectories in the magnetic mass-separation system. The values provided assume an even angular response over a full bin $\Delta \Omega = 22.5^\circ \times 4.5^\circ = 0.031$ sr. However, the true integrated azimuth-elevation response over an IMA sector is roughly half this value. In effect, if the incident ion
distribution is wider than the space angle bin of a sector, half the $G_E$ values shown in Figure 3.8 should be used in equation 3.1 to retrieve the differential flux.

Figure 3.8: Ground-calibrated IMA energy-geometric factor (defined by equation 3.2) for heavy ions in the PAC-4 setting. Note that the values above do not include angular response over a full sector and represent twice the integrated response over an angular bin.

IMA sweeps all 96 steps in the energy table for a single deflector elevation step every 12 s, including ESA voltage settling times between energy steps. The full energy-elevation table is covered every 192 s.

### 3.2.2 Electron Spectrometer (ELS)

In many respects the ASPERA-3 Electron Spectrometer (ELS) is a significantly simpler instrument compared to IMA. Without a deflection system and with no need for mass-separation, the core working ion-optics consist simply of a hemispherical top-hat electrostatic analyzer and an MCP with 16 azimuthal anodes. However, the simpler electrostatic optics also makes ELS more sensitive to stray light. ELS mitigates the photon contamination with a series of EUV-dark baffles and light traps, minimizing stray light inside the instrument, see Figure 3.9.

The ELS electrostatic analyzer resolves a $360^\circ \times 4^\circ$ FOV plane of the electron distribution with a resolution 8% per energy step and orientation dependent on the position of the MU scanner. In the nominal survey mode the energy table sweeps the range 1 eV - 20 keV in 128 logarithmically separated steps with 4 s full-spectrum cadence. The counts recorded by the instrument can be converted to flux using equation 3.2 with analogous factors for ELS. The ELS energy-geometric factor is $G_E = 4.7 \times 10^{-5}$ cm$^2$ sr eV/eV and constant over the full energy range. The integration time for a single ELS energy level is $\Delta t = 28.125$ ms.
Figure 3.9: Hemicylindrical cutaway of the ELS mechanical structure and ion optics. Electrons can enter from all 360° azimuthal angles within a 4° elevation acceptance angle. A series of EUV-dark baffles and light traps mitigates stray light inside the instrument while electrons corresponding to the concurrent ESA voltage/energy setting are guided through the electrostatic analyzer (ESA) and onto a microchannel plate (MCP) with 16 azimuthal sector anodes.
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4. Investigations of ion escape

4.1 From measurements to escape rates

4.1.1 Global distribution functions

Atmospheric ion escape from Mars results from a series of connected processes involving ion production from neutrals and acceleration to escape energy. The orientation of the planet alters the interaction with the solar wind through the strong crustal fields and the interaction incurs changes in the ionosphere as well as vice versa.

Understanding and modeling ion escape from such a complicated and dynamic system is a task fraught with uncertainties. Instead we may directly measure the escaping ion flow as the planet is subjected to naturally varying, though known, upstream conditions. The system response can in turn reveal the dominant escape processes, in effect making the planet into a laboratory. However, individual in-situ measurements may only provide local near-instantaneous single-point distributions, while ion escape is a global and dynamic process, thus many single-point measurements are required to understand the average state of the global plasma flow. The long operation of Mars Express and ASPERA-3 at Mars over more than a solar cycle provides sufficient measurement density and coverage to investigate not only the average ion escape rate, but also the escape response to a wide range of upstream solar wind and solar radiation conditions. IMA and ELS spectra from an example MEX orbit through the deep tail are shown in Figure 4.1. The appearance of the various plasma regions can be seen in the data, including solar wind, magnetosheath, induced magnetotail with several plasma sheet crossings as well as a calm ionosphere with clear $\sim 20$ eV photoelectron peaks.

Individual plasma moments (density, bulk velocity, temperature) are derived from instantaneous (limited by instrument cadence) samples of the local differential flux measurements, and in turn combined to create a global picture of the average plasma environment. Reliable moments require that all significant flux around the spacecraft is observed by the instrument, which can only be guaranteed without a priori assumptions on the absolute distribution if all directions are observed,
Chapter 4. Investigations of ion escape

Figure 4.1: Mars Express ASPERA-3 IMA and ELS raw time-series spectra, integrated over all azimuthal sectors and mass-rings, from an example orbit in June 2017, passing through all major plasma regions at Mars; solar wind, magnetosheath, induced magnetotail and ionosphere. The apparent 192 s modulation in ion spectra is due to the IMA elevation sweeps for energies >50 eV. The relatively energetic ~100 eV electrons in conjunction with similarly accelerated ions during 09:00–09:10 are typical plasma sheet crossings. A mostly calm ionosphere with clear ca. 20 eV photoelectron peaks can be seen around 09:45–10:05.

Instead, over time, many measurements can be combined to cover all angles and locations, however, the measurements are required to be ordered in a suitable reference frame. The structure of the Martian magnetosphere is largely controlled by the draping of the solar wind magnetic field, i.e. by the direction of the local IMF (see section 2.3). Thus it is common to order measurements in the Mars-centered Mars-Sun-Electric field (MSE) reference frame where the direction of the solar wind motional electric field defines the $Z_{\text{MSE}}$-axis, $X_{\text{MSE}}$ points towards the Sun, and $Y_{\text{MSE}}$ follows the right hand orthogonal component of the IMF vector.

Without a magnetometer present on Mars Express, nor on other spacecraft between the failure of MGS (Nov. 2006) and arrival of MAVEN (Sept. 2014), the orientation of the magnetosphere and Mars Express’ position in it remains unknown for the majority of the orbits (see Figure 4.2). Fortunately, the quasi-random orientation of the IMF implies that a cylindrically symmetric approximation in Mars-Sun-Orbit (MSO) coordinates can provide reliable total escape estimates, if sufficient sampling is achieved, an assumption that has been tested and verified in Venus’ induced magnetosphere by Nordström et al. [2013]. In the MSO reference
Figure 4.2: Time coverage of Mars orbiters equipped with plasma and/or magnetometer instrumentation and daily sunspot numbers (SSN) over solar cycles #21–#24 shown in black. 27-day SSN running average shown in red. Mars Express has been operating at Mars for more than a solar cycle, though for most of the operational time there have been no other missions with capabilities to provide contextual magnetic field measurements, forcing the use of a cylindrical coordinate system to organize the measurements.

frame, $X_{MSO}$ is defined by the Sun direction, $Z_{MSO}$ follows the Martian orbit normal and $Y_{MSO}$ completes the right-handed system. An IMA measurement in the instrument’s orthogonal frame

$$j_{IMA} = j_{IMA} \begin{bmatrix} \cos(az) \cos(el) \\ \sin(az) \cos(el) \\ \sin(el) \end{bmatrix}, \quad (4.1)$$

where $az$ and $el$ are the azimuth with respect to $X_{IMA}$, and elevation with respect to the $X_{IMA}$-$Y_{IMA}$ plane, can be converted to MSO coordinates by considering two transformation matrices

$$j_{MSO} = M_{S/C \rightarrow MSO} M_{IMA \rightarrow S/C} j_{IMA}. \quad (4.2)$$

Here, $M_{S/C \rightarrow MSO}$ depends on the orientation and position of the spacecraft, while $M_{IMA \rightarrow S/C}$ relates the instrument reference frame to the MEX spacecraft reference frame.

A cylindrically symmetric 5-dimensional differential flux distribution can be defined, relating to a corresponding phase space density distribution function as

$$j(X, R_{yz}, \theta, \phi, E) = \frac{2E}{m_p^2} f(X, R_{yz}, v_x, v_\theta, v_r) \quad (4.3)$$

where $R_{yz}^2 = Y_{MSO}^2 + Z_{MSO}^2$, $m_p$ is the particle mass and $v_x$, $v_\theta$, $v_r$ are the lateral, tangential and radial velocity components, respectively. $\theta$, $\phi$ and $E$ are the corresponding energy-equivalent spherical coordinates, where $\theta$, $\phi$ are defined relative to the symmetry axis $X_{MSO}$. An illustration of this reference frame is provided.
by Ramstad et al. [2015], Figure 1 in the paper. The 5-dimensional distribution function can be discretized as

\[ j_{i,j,k,l,m} = j(X_i, R_{yz,j}, \theta_k, \phi_l, E_m), \quad (4.4) \]

Note that the index \( j \) differs from the differential flux and the two should not be confused here.

IMA measurements in cylindrical coordinates, \( j_{\text{MSOcyl}} \), may be sorted to the corresponding elements of the discretized distribution function and averaged to yield the average differential flux distribution with scalar elements \( \bar{j}_{i,j,k,l,m} \). The cylindrical components, \( j_{\text{MSOcyl}} \), can be found by a rotation around the intended symmetry axis, \( X_{\text{MSO}} \), to the \( X_{\text{MSO}}-Z_{\text{MSO}} \) plane,

\[
M_{\text{MSO} \rightarrow \text{MSOcyl}} = \begin{bmatrix}
1 & 0 & 0 \\
0 & \cos \theta_{\text{MSO}} & -\sin \theta_{\text{MSO}} \\
0 & \sin \theta_{\text{MSO}} & \cos \theta_{\text{MSO}}
\end{bmatrix}, \quad (4.5)
\]

where the position of the spacecraft in MSO-coordinates determines the rotation angle \( \theta_{\text{MSO}} = -\tan^{-1}(-Y_{\text{MSO}}/Z_{\text{MSO}}) \) and \( j_{\text{MSOcyl}} = M_{\text{MSO} \rightarrow \text{MSOcyl}} \bar{j}_{i,j,k,l,m} \).

Distribution functions are versatile and may be integrated over solid angle and energy to find the net average flux components

\[
\bar{F}_x(X_i, R_{yz,j}) = \sum_k \sum_l \sum_m \bar{j}_{i,j,k,l,m} \cos \theta_k \cos \phi_l \Delta \Omega(\theta_k, \phi_l) \Delta E_m, \quad (4.6a)
\]

\[
\bar{F}_\theta(X_i, R_{yz,j}) = \sum_k \sum_l \sum_m \bar{j}_{i,j,k,l,m} \sin \theta_k \cos \phi_l \Delta \Omega(\theta_k, \phi_l) \Delta E_m, \quad (4.6b)
\]

\[
\bar{F}_r(X_i, R_{yz,j}) = \sum_k \sum_l \sum_m \bar{j}_{i,j,k,l,m} \sin \phi_l \Delta \Omega(\theta_k, \phi_l) \Delta E_m, \quad (4.6c)
\]

where \( \Delta \Omega(\theta_k, \phi_l) = \Delta \theta \Delta \phi \sin \phi_l \) is the solid angle corresponding to the angular element \([\theta_k, \phi_l] \), and the \( \Delta E_m \) factor is the energy step width at energy level \( E_m \). The total flux is simply the magnitude of the components, \( \bar{F} = \sqrt{\bar{F}_x^2 + \bar{F}_\theta^2 + \bar{F}_r^2} \).

With knowledge of the particle mass, \( m_p \), the average density may be computed as the first moment of the distribution function

\[
\bar{n}(X_i, R_{yz,j}) = \sum_k \sum_l \sum_m \bar{j}_{i,j,k,l,m} \Delta \Omega(\theta_k, \phi_l) \sqrt{\frac{m_p}{2E_m}} \Delta E_m. \quad (4.7)
\]

Subsequently, the average velocity components can be found from the net average flux and average density,

\[
\bar{v}_x(X_i, R_{yz,j}) = \frac{1}{\bar{n}(X_i, R_{yz,j})} \sum_k \sum_l \sum_m \bar{j}_{i,j,k,l,m} \cos \theta_k \cos \phi_l \Delta \Omega(\theta_k, \phi_l) \Delta E_m, \quad (4.8a)
\]

\[
\bar{v}_\theta(X_i, R_{yz,j}) = \frac{1}{\bar{n}(X_i, R_{yz,j})} \sum_k \sum_l \sum_m \bar{j}_{i,j,k,l,m} \sin \theta_k \cos \phi_l \Delta \Omega(\theta_k, \phi_l) \Delta E_m, \quad (4.8b)
\]

\[
\bar{v}_r(X_i, R_{yz,j}) = \frac{1}{\bar{n}(X_i, R_{yz,j})} \sum_k \sum_l \sum_m \bar{j}_{i,j,k,l,m} \sin \phi_l \Delta \Omega(\theta_k, \phi_l) \Delta E_m. \quad (4.8c)
\]
4.1 From measurements to escape rates

The choice of appropriate discretization step widths depends on the quantity and resolution of the data that will be used to create the distribution function. The angular resolution of IMA is limited by the separation of the azimuthal sectors, thus a similar angular discretization step $\Delta \theta = \Delta \phi = 22.5^\circ$ is used throughout the papers appended with this thesis, similarly the energy step width $\Delta E_m$ is taken directly from the post-2009 IMA energy table. Examples of global average heavy ion ($O^+$, $O_2^+$, $CO_2^+$) flux, density, and velocity, based on all Mars Express ASPERA-3/IMA PAC-4 data since 2007 are shown in Figure 4.3. The large sample size here allows a fine spatial step size $\Delta X = \Delta R_{yz} = 0.1 R_M$, though the true resolution is limited by the time resolution of the measurements. Here individual 12 s full spectra are used, thus 0.1 $R_M$ corresponds to $\sim 3$ km/s, representative of MEX orbital velocity in the tail. The far tail region ($X_{MSO} \lesssim -2.5 R_M$) is not well covered at the lowest energies <10 eV since MEX last covered this region in 2008.

4.1.2 Escape rates

The majority of the escaping heavy ion flux flows through the tail, as evident in Figure 4.3. The heavy ion pick-up plume can also be seen as smaller radial fluxes outside and near the nominal BS, however, the plume contribution is relatively small ($\sim 20\%$ under high EUV conditions, cf. Dong et al. [2017]) and cannot be accounted for accurately without an estimate of the solar wind motional electric field orientation.

The tail escape rate is calculated from the cylindrically symmetric distribution functions by integrating the downstream flux, $F_\chi$ (eq. 4.6a), in a cross-section of the tail over rings, i.e.

$$ Q(X_i) = -2\pi \Delta R_{yz} \sum_j \bar{F}_\chi(X_i, R_{yz,j}) R_{yz,j}. \quad (4.9) $$

The choice of tail downstream distances, $X_i$, for the cross-section is arbitrary since $Q$ does not depend on $X$ if $X_{MSO} \lesssim -1.0 R_M$ [Nilsson et al., 2012].

4.1.3 Escaping energy flux/escape power

Energy transfer from the solar wind will accelerate ionospheric ions, the total energy flux, i.e. power $P_Q$, of the escaping ions can be calculated, similar to the escape rate, through a cross-section of the tail by integrating the total energy of escaping ions as

$$ P_Q(X_i) = \sum_m E_m q(X_i, E_m) \Delta E_m, \quad (4.10) $$

where $q(X_i, E_m)$ is the differential escape spectrum which can be calculated at any downstream distance, $X_i$, as

$$ q(X_i, E_m) = -2\pi \Delta R_{yz} \sum_j \sum_k \sum_l \bar{f}_{i,j,k,l,m} R_{yz,j} \cos \theta_k \cos \phi_l \Delta \Omega(\theta_k, \phi_l). \quad (4.11) $$

The choice of $X_i$ strongly affects the escape spectrum, and thus $P_Q$, as ions will be accelerated to higher energies the further downstream the cross-section is made, [Nilsson et al., 2012]. The energy of the escaping ions, if above escape energy, is inconsequential for the escape rate, however, the escaping energy flux is useful for quantifying the total power transferred from the solar wind to the escaping ions.
Figure 4.3: Average heavy ion ($O^+$, $O_2^+$, $CO_2^+$) flux, density, and velocity near Mars as calculated through equations 4.6–4.8 from a 5-dimensional distribution function found by integrating all Mars Express ASPERA-3/IMA data taken in PAC-4 mode between 2007–2017. Thin black lines show average velocity direction and thick curves show average Induced Magnetosphere Boundary (IMB) and Bow Shock (BS) locations based on the model by Ramstad et al. [2017a] for upstream solar wind parameters [2 cm$^{-3}$, 400 km/s]. Flux and density grids are here only colored if the local energy-angular coverage is over 75%, though velocity is calculated without coverage requirement. The far tail was last covered in 2008 and thus lacks coverage below 10 eV ($X_{MSO} \lesssim -2.5$ R$_M$). Note that the bulk flow in the tail is slow and almost entirely antisunward behind the planet ($X_{MSO} < 1$), while the dayside flow outside the BS is fast and mainly radial, constituting the oxygen pick-up ion plume.
4.2 Summary and discussion of results

4.2.1 Escape rate dependencies on upstream conditions

Several previous studies have found correlations in the ion escape rate with solar wind dynamic pressure, $p_{\text{dyn}}$, [Lundin et al., 2008a], solar wind flux [Nilsson et al., 2011] and the arrival of solar wind disturbances [Edberg et al., 2010]. These studies have a couple of qualities in common. 1: They are constricted to ion energies above 30 or 50 eV, and in effect mainly include escape through the plasma sheet [Dubinin et al., 2017], which is typically a minority fraction of the total escape rate close to the planet [Nilsson et al., 2012]. 2: They do not simultaneously constrain solar wind and solar radiation properties, which are correlated due to common dependencies on solar cycle and heliocentric distance, thus the influence of one upstream factor may convolute the influence of another.

The large amount of data accumulated by IMA in the subsequent years, with the extended low-energy tables since 2007 ($>10$ eV) and 2009 ($>1$ eV), allows the distribution to be sufficiently sampled while several upstream parameters can be simultaneously constrained. Figure 4.4 shows the effect of upstream solar wind and 1–118 nm photon flux, $F_{\text{XUV}}$, on the escaping heavy ion energy spectra. Increased $F_{\text{XUV}}$ enhances the cold ion outflow while increased $p_{\text{dyn}}$ enhances the plasma sheet escape by accelerating an increasing part of the cold ion outflow population.

![Figure 4.4: Energy spectra of escaping heavy ions (O$^+$, O$_2^+$, CO$_2^+$) in the Martian magnetotail for $-1.9 R_M < X_{\text{MSO}} < -1.5 R_M$. a) upstream solar wind parameters are fixed and 1-118 nm photon flux, $F_{\text{XUV}}$, varies. b) solar wind dynamic pressure varies and $F_{\text{XUV}}$ is fixed. Percentages show the fraction of total escape below 50 eV (indicated by dashed line).](image)

In total, EUV intensity/XUV flux appears to increase the ion escape rate under all solar wind density, $n_{\text{sw}}$, and velocity, $v_{\text{sw}}$, conditions [Ramstad et al., 2015,
Chapter 4. Investigations of ion escape

2017b], a dependency that has also been observed recently through MAVEN observations [Dong et al., 2017]. EUV intensity appears to drive the escape rate through increased photoion production, as indicated by a consistent linear ion escape rate dependence on $F_{\text{XUV}}$, under constrained nominal $n_{\text{sw}}$ and $v_{\text{sw}}$ conditions [Ramstad et al., 2017b]. With EUV intensity constrained, Ramstad et al. [2015] found primarily upstream $n_{\text{sw}}$ to decrease the ion escape rate. A similar influence can be found by partitioning IMA data in the upstream $[p_{\text{dyn}}, F_{\text{XUV}}]$ parameter space, the corresponding dependencies are reported by Ramstad et al. [2017d] as

$$Q(p_{\text{dyn}}, F_{\text{XUV}}) = 10^{24}(0.79 \pm 0.38)p_{\text{dyn}}^{-0.15 \pm 0.06}F_{\text{XUV}}^{0.81 \pm 0.32}$$  \hspace{1cm} (4.12)

where $p_{\text{dyn}}$ and $F_{\text{XUV}}$ are defined in units of [nPa] and $[10^{14} \text{ m}^{-2} \text{ s}^{-1}]$, respectively. The function is plotted in Figure 4.5. Qualitatively, the ion escape rate has a weak inverse dependence on $p_{\text{dyn}}$ and an approximately linear dependence on $F_{\text{XUV}}$.

![Figure 4.5:](image)

Figure 4.5: The ion escape rate as a function of upstream solar wind dynamic pressure and 1-118 nm photon flux (eq. 4.12). Black dots with standard error bars represent escape rates derived from global average distribution functions. Red areas represent the corresponding 1σ and 1.96σ confidence regions. Adapted from Ramstad et al. [2017d].

A fast and extremely dense [39 cm$^{-3}$, 730 km/s] CME arrived at Mars in July 2011, providing an extreme case to study the solar wind influence on the ion escape rate. The event was analyzed by Ramstad et al. [2017c] and found to increase tailward heavy ion fluxes to $\times 3$ of average, though to have no significant effect on the escape rate due to the large compression of the tail region. Instead, the solar wind accelerated ions to ca. $\times 20$ higher energies, indicating a much higher energy transfer, despite no significant effect on the escape rate.
4.2 Summary and discussion of results

4.2.2 Solar wind - ion escape coupling

The increase in energy flux during the July 2011 event is consistent with the trends observed at conditions closer to nominal. The total energy flux of the tailward escaping ions, \( P_Q \) (eq. 4.10), increases with increased available upstream solar wind power,

\[
P_{sw} = \frac{\rho_{sw} v_{sw}^3}{2} A_{IMB},
\]

(4.13)

where \( A_{IMB} \) is the cross-section of the obstacle to the solar wind, i.e. the IMB. However, \( P_Q \) increases at less than a proportional rate of change since the energy transfer, i.e. coupling, efficiency

\[
k = \frac{P_Q}{P_{sw}},
\]

(4.14)

decreases with upstream \( p_{dyn} \) at a rate \( k \propto p_{dyn}^{-0.74 \pm 0.13} \). Overall, the induced magnetosphere provides efficient protection for the ionosphere contained inside the IMB, as \( \lesssim 2\% \) of the power available in the solar wind is transferred inside the tail near the planet [Ramstad et al., 2017b], and a smaller fraction for high \( p_{dyn} \), see Figure 4.6.

![Figure 4.6: Solar wind/ion escape coupling efficiency as a function of solar wind dynamic pressure for high (red) and low (black) EUV conditions, showing similar trends for both constraints. Adapted from Ramstad et al. [2017b], with permission.](image)

4.2.3 Role of the crustal fields

The mostly southern hemisphere fields of remnant crustal magnetization and their effect on the solar wind interaction were introduced in Chapter 2. A flux rope
observed by MGS in the vicinity of the crustal fields was estimated by Brain et al. [2010] to channel \((0.08 – 1.6) \times 10^{24}\) ions/s, assuming charge neutrality with the observed electrons, i.e. potentially comparable to the total global escape rate. However, the flux ropes are not always apparent near the crustal fields, thus the total estimated contribution to global escape is smaller (~5–10%). A small effect is supported by tail observations by Nilsson et al. [2011], who found 38% ± 11% of escape outflowing from the southern quadrant, compared to the north–south total. In order to avoid bias from observing different outflow regions under different upstream conditions, Ramstad et al. [2016] calculated the ion escape rate for the northern and southern geographical hemispheres separately while also constraining upstream solar wind and EUV conditions to nominal intervals \([1–3 \text{ cm}^{-3}, 350–450 \text{ km/s}, 4–5 \text{ mW/m}^2]\). The result was a smaller difference \((46% ± 18%)\) compared to Nilsson et al. [2011], showing that the overall effect of the crustal fields on ion escape is negligible.

In addition, Ramstad et al. [2016] also studied the effect of crustal field orientation by comparing the global ion escape rate for varying SZAs of the strongest crustal field region \([178°E, 53°S]\). They found a lowered escape rate for low dayside SZA \((28°–60°)\), an enhancement in cold ion outflow for high dayside SZA \((60°–80°)\) and no mutual difference between terminator and night-side SZA. The cold ion outflow enhancement at SZA = 60° is closely matched by a sharp increase in the electron scale height over areas where the crustal field strength dominates the induced field strength [Andrews et al., 2015], see the comparison in Figure 4.7. The consistent enhancement indicates a significantly different interaction with the solar wind dependent on SZA. We may consider a tentative explanation: Some of the crustal fields are known to be sufficiently strong to affect the compressed solar wind flow in the sheath [Crider, 2004; Wang et al., 2015], and this effect can be expected to be particularly strong at high SZA where the local solar wind dynamic pressure is low. Here, the intrinsic crustal fields interact directly with the turbulent sheath. If the interaction is similar to solar wind interaction with larger magnetospheres we may expect significant wave heating at lower altitudes through primarily Alfvén wave propagation along the field lines, which connect the sheath interaction to all ionospheric layers. The heated ionospheric plasma may subsequently outflow in the form of the flux tubes observed by Brain et al. [2010]. Below 60° SZA, the crustal fields may simply compress under the higher local dynamic pressure and, in combination with a more hot, though slower, local sheath flow have less of an obstacle effect, rather simply increasing the stand-off distance between the ionosphere and the solar wind. Close to 90° SZA, photoion production is low and thus little ionospheric plasma is produced that may outflow.

### 4.2.4 Total atmosphere escaped as ions

The total amount of atmospheric mass removed through ion escape can be estimated with an accurate understanding of the system response to the influential parameters. Ionizing photon flux produces the ions available for escape and the solar wind provides the necessary energy to reach escape velocity and influences transport by altering the strength and morphology of the induced magnetosphere. Under the assumption that \(p_{\text{dyn}}\) and \(F_{\text{XUV}}\) are the only significant factors, Ramstad et al. [2016]...
4.2 Summary and discussion of results

Figure 4.7: Influence of crustal field orientation on ion escape rate. Upper panel shows the ion escape rate for varying SZA of the strongest crustal field region [178°E, 53°S], as measured by IMA in the tail. Bottom panels show the plasma density and scale height against altitude and SZA in areas where the crustal field strength is higher than a threshold value related to the strength of the draped magnetic field, as measured by the MARSIS radar on MEX. Note the increase in plasma scale height and ion escape at $SZA = 60^\circ$. Upper and lower panels are adapted, with permission, from Ramstad et al. [2016] and Andrews et al. [2015], respectively.

[2017d] use the empirical escape model from equation 4.12, informed by modeled evolution of solar wind and EUV from Ribas et al. [2005], Wood [2006] and Airapetian et al. [2016] to estimate the ion escape rate through time. The integrated escape rate accounts for an equivalent $4.8 \pm 1.1$ mbar lost since the mid-late Hesperian (3.3 Ga), from roughly which time the last pieces of evidence for an active Martian hydrosphere date. While the contribution from the energetic oxygen pick-up ion plume is only roughly included and is insensitive to EUV irradiance, possibly due to an increased IMB altitude with EUV [Dong et al., 2017; Ramstad et al., 2017a], the plume is taken to provide an upper estimate of the total atmosphere lost as $\sim 6$ mbar, under the assumption that the increase in the plume escape rate may completely compensate the tail ion escape rate inverse dependence on $p_{\text{dyn}}$.

Extending the extrapolation to the late Noachian, from which remnant terraines
feature ubiquitous evidence for long-standing lacustrine environments and precipitation (see Chapter 1), only yields $6.3 \pm 1.9 \text{ mbar}$ lost since, and an analogous upper estimate $\sim 9 \text{ mbar}$ [Ramstad et al., 2017d]. The extrapolation through time, including escape rate and the upstream parameters, is shown in Figure 4.8.

---

**Figure 4.8**: Extrapolation of (a)–d) upstream solar wind parameters [Wood, 2006; Airapetian et al., 2016], e) 1–118 nm photon flux [Ribas et al., 2005], f) corresponding ion escape rate based on eq. 4.12, and g) integrated equivalent surface pressure lost through ion escape since any point in time back to 3.9 Ga. The blue areas mark $1\sigma$ and $1.96\sigma$ confidence intervals for the mean. The gray shaded area in panel g) roughly indicates the timing of the youngest traces of precipitation on the Martian surface and thus a Martian hydrological cycle ($\sim 3.3 \text{ Ga}$). The extrapolated ion escape trends found can only account for $4.8 \pm 1.1 \text{ mbar}$ lost since. Extended to the late Noachian (3.9 Ga), ion escape still can only account for $6.3 \pm 1.9 \text{ mbar}$. Adapted from Ramstad et al. [2017d].
4.3 Conclusions and implications

4.3.1 Influence of ion escape on the evolution of Mars’ atmosphere

The strong solar wind and radiation environment around the younger Sun has eroded an equivalent surface pressure $\lesssim 6$ mbar through ion escape since the mid-late Hesperian (3.3 Ga), thus ion escape enhancement by upstream factors alone cannot explain the collapse of the early Martian hydrosphere. Extended to the late Noachian, ion escape still cannot account for significantly more than ca. 10 mbar, as such, the ion escape channel appears to have had a very small influence on the evolution of the Martian atmosphere [Ramstad et al., 2017d].

4.3.2 Induced magnetosphere screening of the ionosphere

Typically less than 1% of the power available in the solar wind is transferred to escaping ions inside the IMB [Ramstad et al., 2017b], and while a higher dynamic pressure accelerates ions to higher energies, the total escape rate is not strongly affected [Ramstad et al., 2015, 2017b,c]. This suggests that the power transferred from the solar wind on the dayside is mainly deposited in the plasma of the topside ionosphere, while the bulk of the ionosphere is located and protected at lower altitudes.

The exact mechanism that leads to an apparent lack of energy-limitation to ion escape is unknown. A significant potential difference up to 6.5 V may be present between the ionosphere and the far tail [Collinson et al., 2015, 2017], which on its own would be sufficient to accelerate the ionospheric plasma above escape velocity as polar outflow. Subsequently, the already escaping plasma would be accelerated further through $\mathbf{J} \times \mathbf{B}$-acceleration in the plasma sheet Ramstad et al. [2017b], increasing escaping ion energy flux, though not the ion escape rate, as observed.

4.3.3 Crustal magnetic fields

The crustal magnetic fields do not significantly affect the average ion escape rate, at least not under nominal solar wind and XUV conditions. However, the influence of the crustal fields appears to depend strongly on their orientation relative the solar wind flow [Ramstad et al., 2016], indicating a more nuanced picture of the general role of intrinsic magnetic fields in ion escape processes. The crustal magnetic fields apparently do not always shield the ionosphere, rather the opposite, increasing coupling between the solar wind and the ionosphere at high dayside SZA orientations.

4.4 Discussion

As often is the case in science, providing an answer to a question leads to several more raised. The studies undertaken through this thesis mainly focus on the ion escape rate, partly due to the absence of field measurements on MEX, which is required to investigate many of the potential underlying physical processes. However, the simultaneous presence of Mars Express and MAVEN may provide the means to answer some questions in future investigations.
4.4.1 What process limits the ion escape rate?
A decreasing escape rate with increased solar wind $n_{sw}$ or $p_{dyn}$ indicates that the Martian escape rate is not limited by energy transfer from the solar wind, instead, the positive near-linear dependence on $F_{XUV}$ rather indicates photoion production rate to be the limiting factor. However, globally, ions are produced at rates several orders of magnitude higher than ions are escaping, which suggests that the source region is relatively thin and that diffusive transport from the dense thermal ionosphere is limited, perhaps due to magnetization of the top-side ionosphere, which increases with upstream $p_{dyn}$. Identifying the source region and solar wind energy deposition altitudes will likely reveal the factors that limit the ion escape rate and give rise to the observed trends.

4.4.2 Does a thicker atmosphere affect ion escape?
Naturally, in-situ measurements of escape processes can only give information on escape from the modern Martian atmosphere. The influence of total atmospheric mass content on ion escape is unknown and, while ion escape rates are low, a more massive atmosphere, removed by other escape channels or sequestration processes, could hypothetically affect ion escape rates. Mars is uniquely suited for studying the effects of atmospheric mass content under similar gravity and upstream conditions due to the seasonal CO$_2$ polar sublimation/deposition cycle. However, to detect the effects of atmospheric mass on ion escape will require constraints on all other influential factors, including solar wind parameters, XUV flux, and crustal field orientation, requiring a statistical base of measurements likely larger than presently available from any currently active mission.

4.4.3 Constraining the plume ion escape channel solar wind dependence
The energetic oxygen ion pick-up plume is a relatively minor escape channel contributing 20%–30% of the total escape rate, and a smaller percentage at high EUV irradiance due to the insensitivity of this channel [Dong et al., 2017]. Ramstad et al. [2017d] assumes as an upper estimate that the plume ion escape rate increases sufficiently to completely balance the decrease in the tail escape rate, however, the plume ion escape rate dependence on solar wind properties is as of yet not investigated and should be constrained.

4.4.4 What drove the high ion escape rates observed in 1989?
In 1989 the Phobos-2 orbiter measured escape rates of $(2−3) \times 10^{25} \text{ s}^{-1}$ [Lundin et al., 1990; Ramstad et al., 2013], i.e. significantly higher compared to the recent estimates by Mars Express and MAVEN. While solar activity was higher in 1989, the XUV fluxes as estimated from the FISM-M model by Thiemann et al. [2017] can only account for $\sim 6 \times 10^{24} \text{ s}^{-1}$, given the dependence from Ramstad et al. [2017b]. The fluxes observed by Phobos-2 in the tail do not correlate well with upstream solar wind parameters on an orbit-by-orbit basis, implying influence by an unknown driver. An intense SEP environment can be expected under such a strong cycle and the influence of SEPs on ion escape remains perhaps the least explored potential upstream driver. However, the Phobos-2 escape rate was obtained by
extrapolating measurements from only 18 tail passes and thus a factor 2-3 may not be surprising. Another possibility is that cross-talk between heavy ion species in the Phobos-2/ASPERA Wien filter was more severe than thought, artificially inflating total fluxes as fluxes of individual species were added.

4.4.5 Do intrinsic magnetospheres protect planets from solar wind?

Ion escape from Mars is but one manifestation of the physical processes that govern ion escape from all planets. All Martian ion escape rates, across all upstream conditions, are in the range \((1.5 - 6) \times 10^{24} \text{ s}^{-1}\) as measured by MEX, and insensitive to solar wind parameters and energy transfer. In contrast, increased energy transfer is strongly coupled to increased ion escape rates from the Earth, which thus range widely from ca. \(10^{25} \text{ s}^{-1}\) to over \(10^{26} \text{ s}^{-1}\) [Li et al., 2017; Slapak et al., 2017], i.e. an order of magnitude or higher compared to the Martian ion escape rate and enhanced with both intensified solar wind, IMF as well as EUV irradiance of the polar cap, indicating an ion escape process for heavy magnetized planets limited in both energy transfer and production simultaneously, perhaps at different altitudes.

Earth’s magnetosphere deflects the bulk of the solar wind at large distances upstream from the planet, though it also provides a large cross-section area for energy transfer. Moreover, the vertical field-lines connect all layers of the ionosphere to the solar wind, providing a channel for energy transfer through Alfvén waves and polarization electric fields.

As part of the greater picture, the ion escape drivers at Venus are currently relatively unexplored. With a gravity well similar to Earth and an induced magnetosphere much more like that of Mars, Venus constitutes a missing middle piece in the ion escape puzzle. In contrast to Mars and Earth, ion escape at Venus decreases with increased EUV irradiance [Kollmann et al., 2016], indicating a strongly energy-limited ion escape process [Ramstad et al., 2017b].


### List of acronyms

<table>
<thead>
<tr>
<th>Acronym</th>
<th>Definition</th>
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<tbody>
<tr>
<td>ASPERA</td>
<td>Automatic Space Plasma Experiment with a Rotating Analyzer (Phobos-2)</td>
</tr>
<tr>
<td>ASPERA-3</td>
<td>Analyzer of Space Plasmas and Energetic Atoms (Mars Express)</td>
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<tr>
<td>BS</td>
<td>Bow Shock</td>
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<tr>
<td>CIR</td>
<td>Corotating Interaction Region</td>
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<tr>
<td>CME</td>
<td>Coronal Mass Ejection</td>
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<tr>
<td>CTX</td>
<td>Context Camera</td>
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<tr>
<td>ENA</td>
<td>Energetic Neutral Atom</td>
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<tr>
<td>ESA</td>
<td>Electrostatic Energy Analyzer</td>
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<tr>
<td>ESA</td>
<td>European Space Agency</td>
</tr>
<tr>
<td>ELS</td>
<td>Electron Spectrometer</td>
</tr>
<tr>
<td>EUV</td>
<td>Extreme Ultraviolet</td>
</tr>
<tr>
<td>FOV</td>
<td>Field-of-View</td>
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<tr>
<td>FWHM</td>
<td>Full Width at Half Maximum</td>
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<tr>
<td>HGA</td>
<td>High-Gain Antenna</td>
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<tr>
<td>HiRISE</td>
<td>High Resolution Imaging Experiment</td>
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<tr>
<td>IB</td>
<td>Ionosphere Boundary</td>
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<tr>
<td>IMA</td>
<td>Ion Mass Analyzer</td>
</tr>
<tr>
<td>IMB</td>
<td>Induced Magnetosphere Boundary</td>
</tr>
<tr>
<td>IRF</td>
<td>Institutet för rymdfysik (Swedish Institute of Space Physics)</td>
</tr>
<tr>
<td>MAG/ER</td>
<td>Magnetometer/Electron Reflectometer</td>
</tr>
<tr>
<td>MAVEN</td>
<td>Mars Atmosphere and Volatile EvolutioN</td>
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<td>MCP</td>
<td>Micro-channel plate</td>
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<td>MEX</td>
<td>Mars Express</td>
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<td>MGS</td>
<td>Mars Global Surveyor</td>
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<td>MLI</td>
<td>Multi-Layer Insulation</td>
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<tr>
<td>MOI</td>
<td>Mars Orbit Insertion</td>
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<tr>
<td>MOLA</td>
<td>Mars Orbital Laser Altimeter</td>
</tr>
<tr>
<td>MRO</td>
<td>Mars Reconnaissance Orbiter</td>
</tr>
<tr>
<td>MSO</td>
<td>Mars-Sun-Orbit</td>
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<td>MSE</td>
<td>Mars-Sun-Electric field</td>
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<tr>
<td>MSL</td>
<td>Mars Science Laboratory</td>
</tr>
<tr>
<td>MU</td>
<td>Main Unit</td>
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<tr>
<td>NASA</td>
<td>National Aeronautics and Space Administration</td>
</tr>
<tr>
<td>NPD</td>
<td>Neutral Particle Detector</td>
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<tr>
<td>NPI</td>
<td>Neutral Particle Imager</td>
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<tr>
<td>S/C</td>
<td>Spacecraft</td>
</tr>
<tr>
<td>SEE</td>
<td>Solar Extreme Ultraviolet Explorer</td>
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<tr>
<td>SEP</td>
<td>Solar Energetic Particle</td>
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<tr>
<td>SSN</td>
<td>Sunspot number</td>
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<tr>
<td>SZA</td>
<td>Solar-zenith angle</td>
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<tr>
<td>TIMED</td>
<td>Thermosphere Ionosphere Mesosphere Energetics and Dynamics</td>
</tr>
<tr>
<td>TBTV</td>
<td>Thermal Balance/Thermal Vacuum</td>
</tr>
<tr>
<td>XUV</td>
<td>X-ray and extreme Ultraviolet</td>
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