# 7

# Climates of terrestrial planets

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*Suppose we detect a planet half the size of Venus orbiting a 5 billion year old M-type star at 0.5 AU. To our surprise the planet has detectable radiation belts. How might the planet's climate and surface habitability differ from that of Venus?*

The prospect that the scientific community might be faced in the next decade or two with questions like the one above is exciting. It is also daunting, because we will likely be required to make inferences about distant planets based on partial information about their environment, orbit, and characteristics. Fortunately, we have at our disposal abundant information about the planets in our own solar system, and more than a half century of practice studying how very complex climate systems function.

This chapter provides the interested reader with an overview of the likely links between heliophysics and climate. Climate is typically defined as the long-term (multi-decade or longer) average of weather. For example, Merriam Webster defines climate as "the average course or condition of the weather at a place usually over a period of years as exhibited by temperature, wind velocity, and precipitation", while Wikipedia currently describes climate as "a measure of the average pattern of variation in temperature, humidity, atmospheric pressure, wind, precipitation", atmospheric particle count, and other meteorological variables in a given region over long periods of time. Climate studies therefore investigate properties of planetary atmospheres – properties that are influenced by both intrinsic characteristics of the planet and by interactions with the host star. To determine whether a planet is or has been habitable at its surface, therefore, it is helpful to understand how a variety of processes act together to influence climate over time. Heliophysical processes are important components of this understanding.

## 7.1 Current climates of terrestrial planets

This chapter focuses on the global climates of terrestrial (rocky) planets Venus, Earth, and Mars. Because they have atmospheres, gas and ice giant planets such

Table 7.1 *Present characteristics and climates of the terrestrial planets*

| | Venus | Earth | Mars |
|---|---|---|---|
| Radius | 6050 km | 6400 km | 3400 km |
| Heliocentr. dist. | 0.72 AU | 1 AU | 1.52 AU |
| Rot. period | 243 days | 24 hours | 24.6 hours |
| Surface temp. | 740 K | 288 K | 210 K |
| Surface press. | 92 bar | 1 bar | 7 mbar |
| Composition | 96% $CO_2$; 3.5% $N_2$ | 78% $N_2$; 21% $O_2$ | 95% $CO_2$; 2.7% $N_2$ |
| $H_2O$ content | 20 ppm | 10,000 ppm | 210 ppm |
| Precipitation | None at surface | Rain, frost, snow | Frost |
| Circulation | 1 cell/hemisph., quiet at surface but very active aloft | 3 cells/hemisph., local and regional storms | 1 cell/hemisph. or patchy circulation, global dust storms |
| Max. surf. wind | $\sim$3 m/s | >100 m/s | $\sim$30 m/s |
| Seasons | None | Comparable north. and south. seasons | Southern summer more extreme |

as Jupiter, Saturn, Uranus, and Neptune have climates, as do planetary moons with gravitationally bound atmospheres ranging from very thick (e.g., Saturn's moon Titan) to considerably more tenuous (e.g., Jupiter's moons Io and Europa). Dwarf planets (e.g., Pluto) also have climates, though we know relatively little about them at present. The terrestrial planets are of special interest because they are thought to have been habitable at their surfaces at some point during solar system history. They formed under similar conditions (see Ch. 4 in Vol. III), with early atmospheres that were more similar than they are today. The present day climates of Venus and Mars provide a useful contrast to that of Earth, and exploration of the root causes for differences in the present climates of all three planets allows us to better understand the processes that control climate everywhere. Their current climates are summarized in Table 7.1, and discussed briefly below.

The surface pressures of the three planets differ by more than four orders of magnitude, with a Martian pressure less than 1% that of Earth, and Earth's pressure a little more than 1% that of Venus. The atmospheric density near the surface of Venus is approximately 8% that of liquid water, while the atmospheric density near the surface of Mars is comparable to the density at altitudes higher than $\sim$35 km on Earth, more than three times the altitude of Mt. Everest. Despite their large differences in mass, the atmospheres of Venus and Mars have similar bulk compositions, with carbon dioxide ($CO_2$) comprising $\sim$95% by volume, followed by molecular nitrogen ($\sim$3%), and argon ($\sim$1%). Earth's atmosphere, by contrast, is composed mainly of nitrogen and oxygen, followed by argon. Earth's atmospheric composition likely mirrored that of Venus and Mars early on, but much of Earth's atmospheric $CO_2$ now resides in carbonates on the ocean floors, leaving nitrogen as the most common constituent. Earth's abundant atmospheric oxygen is believed to have been contributed by photosynthetic bacteria.

The surface temperatures of the three planets also differ widely, in part due to the distance of each planet from the Sun and in part due to the quantity of greenhouse gases in each atmosphere. Earth is the only of the three planets with a surface temperature (and pressure) appropriate for liquid water to be stable for long periods of time, thanks to $\sim$30 K of greenhouse warming. The atmosphere of Venus is too hot for water to exist as liquid at the surface, while the Martian atmosphere has too low a surface pressure (liquid water would sublime, except at the lowest elevations). The atmosphere of Venus is very dry, indicating that any surface water driven into the atmosphere by the high temperatures no longer resides there. The relative atmospheric water content at Mars is an order of magnitude larger than at Venus and, given the low atmospheric pressure, is often nearly saturated. Despite the near 100% Martian relative humidity, Earth still has roughly 50 times more water molecules (per particle of atmosphere) than Mars. The composition, temperature, and water content lead to different forms of precipitation on the three planets. Earth has a variety of forms of water precipitation, while Mars has carbon dioxide and water frost. Venus has no precipitation at the surface due to its high temperatures; any precipitation that forms higher in the atmosphere would turn to vapor before reaching the ground.

Circulation patterns on the three planets also differ. Earth possesses three circulation cells in each hemisphere, leading to prevailing winds organized by latitude. The circulation results, in a simplified sense, from an equator-to-pole temperature gradient that causes warm air to rise at the equator and fall at the poles. Earth's rotation provides a Coriolis influence that breaks the circulation cells into three regions, keeping the warmest air relatively confined at low latitudes. Venus, by contrast, rotates very slowly. Thus, heat is transferred efficiently from the equator to polar regions, leading to uniform surface temperatures as a function of latitude and local time (see Bullock and Grinspoon, 2013). Mars rotates at nearly the same rate as Earth but has only one circulation cell per hemisphere, though there are some arguments to suggest that while there is a net circulation, air tends to move in localized regional cells (see Rafkin *et al.*, 2013). Air at the surface of Mars moves sufficiently quickly to drive dust devil activity, while the surface of Venus is very still. At higher altitudes on Venus, however, the atmosphere superrotates on time scales of days (e.g., Kouyama *et al.*, 2013).

While Earth's seasonal variations, caused by a 23.5° tilt relative to its orbital plane, will be well known to the reader, seasonal variations on Venus and Mars are substantially different. Venus has nearly no seasonal variation due to a very small ($\sim$3°) axis tilt. Mars has a tilt of 25°, similar to that of Earth, but the planet's greater orbital eccentricity (a 21% difference between the perihelion and aphelion distances compared to 1.4% and 3.3% for Venus and Earth, respectively) leads to shorter and more intense summers in the southern hemisphere compared to the north. Strong heating during southern summer drives enhanced dust devil activity, which can couple across circulation cell boundaries and grow into planet-encompassing dust storms that last several weeks.

## 7.2 Evidence for climate change

Abundant evidence points to changes in the climate of all three terrestrial planets on a variety of time scales. Here, we focus on evidence for climate change over tens of thousands of years or longer. The reader is also directed to discussions in Chs. 4, 11, and 12 in Vol. III (cf., Table 1.2).

The most compelling evidence for climate change on Venus comes from measurements of the isotopes deuterium and hydrogen in the atmosphere today (Fig. 7.1a). Deuterium is far scarcer than hydrogen in the atmospheres of all planets. However, the ratio of deuterium to hydrogen (D/H) in the Venus atmosphere – about two deuterium atoms for every 100 hydrogen atoms – is more

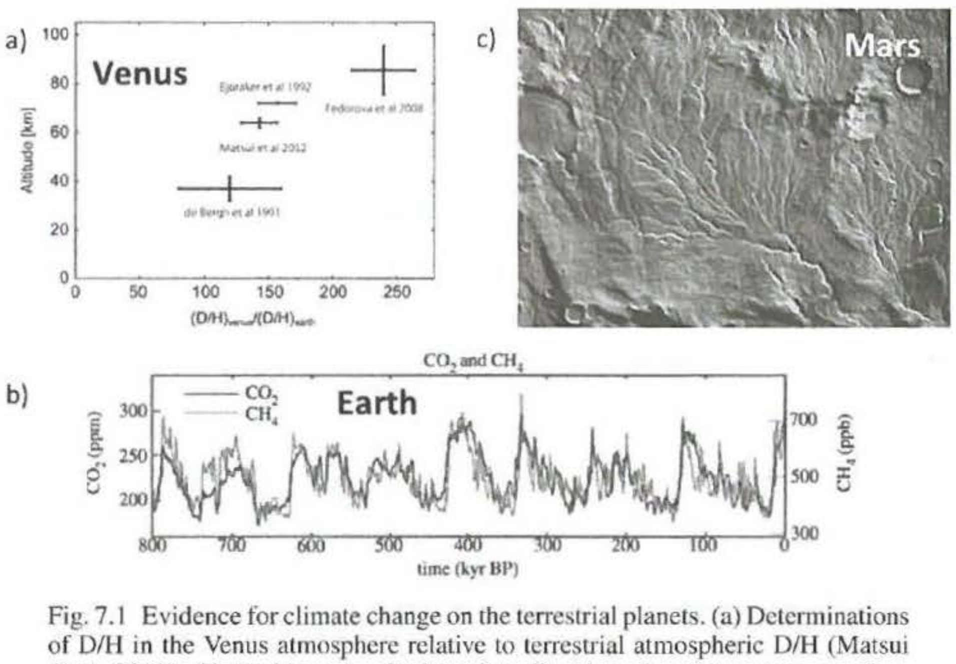


Fig. 7.1 Evidence for climate change on the terrestrial planets. (a) Determinations of D/H in the Venus atmosphere relative to terrestrial atmospheric D/H (Matsui *et al.*, 2012); (b) Earth's atmospheric carbon dioxide and methane concentrations as a function of time, as determined from ice cores (Hansen *et al.*, 2013); (c) a dendritic river valley network in the Warrego Valles region of Mars (courtesy NASA Viking).

than 100 times the same ratio calculated for Earth (Donahue *et al.*, 1982) and most other solar-system objects. There is little reason to expect that Venus formed with a D/H ratio significantly different from that of Earth, so we infer that the D/H ratio on Venus increased after the planet formed. Specifically, it is thought that hydrogen atoms (possibly from a primordial ocean) preferentially escaped the planet's gravity compared to deuterium and were lost to space. Section 7.5 explains this physical process in detail. The loss of hydrogen to space at Venus simultaneously explains the measured D/H ratio and the relative scarcity of water in the Venus atmosphere today: water was dissociated in the atmosphere and the hydrogen removed to space.

One might wonder whether the surface of Venus holds any clues about its past climate, in the same way that the geologic records of Earth or Mars can teach us about time periods billions of years ago. Unfortunately, the low-impact crater abundance at Venus suggests that the planet has been entirely resurfaced within the past several hundred million years. Whether this occurred in a global event or gradually over time is debated (Phillips *et al.*, 1992; Schaber *et al.*, 1992), but the implications for inferring climate change are identical in either case: it is not straightforward to use surface features as indicators of ancient climatic conditions at Venus.

In contrast to Venus, evidence for climate change on Earth is abundant and comes in many different forms. On time scales of hundreds to thousands of years, the measured growth rate of tree rings and coral as a function of time are used to infer various aspects of climate such as atmospheric and ocean temperature, precipitation, and ocean salinity. On time scales extending back hundreds of thousands of years, layered ice cores are used to infer atmospheric temperatures (through isotope ratios in the ice and layer thickness), atmospheric composition (through gases trapped in air pockets in the ice), and even which plants were present (through trapped pollen). Beyond a million years ago we are reliant on rock geochemistry, fossils, and sediment to provide information about atmospheric temperature, composition, climate shifts, and even surface pressure.

The terrestrial climate record from all of these sources suggests that Earth's climate varies on many time scales, with departures in temperature of as much as 10–15 °C over Earth's history (e.g., Hansen *et al.*, 2013). There are many inferred cold (glaciation) and warm periods that have been tied with changes in atmospheric conditions and diversity of life. Similarly, there are a few major changes in atmospheric composition, the most notable of which is the oxygenation of the terrestrial atmosphere more than two billion years ago (Bekker *et al.*, 2004; Ch. 4 in Vol. III), likely caused by the rise of oxygen-producing bacteria and the subsequent depletion of sinks for oxygen at Earth's surface (e.g., Kaufman *et al.*, 2007). Analysis of the size and depth of fossilized raindrop imprints in sedimentary rock even suggests that Earth's surface pressure has varied by as much as a factor of two over 2.7 billion years (Som *et al.*, 2012). Taken together, the evidence provides a caution against interpreting the present day climates of other terrestrial planets too finely, and assuming only monotonic changes in planetary climates over billion year time scales. At the same time, one of the most notable aspects of the terrestrial record is the fact that water has existed as liquid at the surface for most of the planet's history, suggesting that despite short-term deviations Earth's climate has been relatively stable over its history, in likely contrast to Venus and Mars.

Mars also provides several lines of evidence suggesting past climate that differs from today. This evidence can be broadly classified as geomorphologic, geochemical, or atmospheric (Jakosky and Phillips, 2001). Geomorphologic evidence includes surface features that are unlikely to have formed in today's environment. These include dry dendritic (branching) river valley networks (Fig. 7.1c), river delta deposits, possible regions of sedimentary rock, smoothed and rounded rocks imaged by Mars rovers, and possible ancient ocean shorelines. These features all suggest an ancient Mars where liquid water was abundant and active in shaping the surface of the planet. Further, highly eroded crater rims and a paucity of small craters relative to what might be expected from the abundance of large craters suggest that the ancient atmosphere was much more efficient at eroding surface features (i.e., thicker) than today – perhaps as thick as 0.5–3 bar, or even more.

Geochemical evidence on Mars demonstrates that water was the liquid responsible for creating the observed surface features, and not some other chemical species. Observations made from both orbiting spacecraft and surface rovers show the presence on both regional and local scales of minerals that require water to form and/or incorporate water as part of their crystal structure. These minerals include sheet silicates (phyllosilicates, including clays), sulfates, and carbonates.

Finally, a number of Martian atmospheric isotope ratios (D/H, $^{38}Ar/^{36}Ar$, $^{13}C/^{12}C$, $^{15}N/^{14}N$, $^{18}O/^{16}O$) point to the stripping of atmospheric particles to space over billions of years, similar to the inference drawn from D/H measurements at Venus (Jakosky and Phillips, 2001). Each of these species is enriched in the more massive isotope, suggesting that the lighter isotope has been preferentially removed to space from the upper atmosphere. In some instances (e.g., argon), no other processes are known that are capable of altering the isotope ratio. Together, the isotope ratios suggest that 50%–90% of the total atmospheric content has been removed to space from stripping processes alone.

## 7.3 How do climates change?

With abundant evidence that planetary climates are not static, we next turn our attention to the planetary characteristics and processes that can be responsible for changing climate. For this discussion we focus primarily on surface temperature, which directly or indirectly influences many other aspects of climate. The reader is also directed to Vol. III, Ch. 16 (Brasseur *et al.*, 2010).

Surface temperature is determined by the global energy budget for the atmosphere. An expression for the energy budget is given by

$$\frac{S}{d^2}(1 - A)\pi R_{\mathrm{p}}^2 = \sigma T_{\mathrm{eff}}^4 4\pi R_{\mathrm{p}}^2, \tag{7.1}$$

where $S$ is the stellar irradiance at 1 AU ($\sim$1360 W/m$^2$ for our Sun, sometimes called the solar constant), $d$ is the distance in units of AU from the star to the planet, $A$ is the planet's Bond albedo (0 for perfectly absorbing and 1 for perfectly reflective), $R_{\mathrm{p}}$ is the radius of the planet, $\sigma$ is the Stefan–Boltzmann constant, and $T_{\mathrm{eff}}$ is the effective radiating temperature of the planet. This equation assumes (quite reasonably) that incident and outgoing radiation at a planet are balanced, and that the planet is rotating (i.e., effectively redistributes energy). On the left-hand side, the incident flux of energy from the star at the location of the planet ($S/d^2$) strikes the disk cross section of the planet facing the star ($\pi R_{\mathrm{p}}^2$) and is absorbed at the surface of planet according to its albedo ($1 - A$). On the right-hand side, the planet radiates energy away ($\sigma T_{\mathrm{eff}}^4$) over its entire surface area ($4\pi R_{\mathrm{p}}^2$). For the solar system Eq. (7.1) can be reduced to $T_{\mathrm{eff}} = 280(1 - A)^{1/4} d^{-1/2}$ K.

The temperature in Eq. (7.1) is effective temperature, which is the temperature of a black body emitting the same amount of radiation to space as the planet. Effective temperature can be related to surface temperature of a planet with an atmosphere under a few assumptions. Here, it is sufficient to write

$$T_s^4 = (1 + \tau) T_{\mathrm{eff}}^4, \tag{7.2}$$

where $T_s$ is the surface temperature and $\tau$ is the optical depth of the atmosphere. This expression is derived under the assumption of a plane-parallel gray atmosphere (i.e., assumed to be wavelength independent) under radiative equilibrium. The atmosphere consists of $\tau$ slabs of unit optical depth, each of which absorbs all radiation emitted from adjacent layers, and emits black-body radiation only to adjacent layers.

There are a few things to note from Eqs. (7.1) and (7.2). First, the size of a planet plays no role in the global energy balance. On average, each portion of a planetary surface absorbs and radiates its fair share of energy, so that the size of a planet is unimportant. Second, the interior heat from a planet plays essentially no role in radiative equilibrium, as evidenced by the fact that only effective and surface temperature appear in Eqs. (7.1) and (7.2), and not the temperature of the planetary interior. This is true for terrestrial planets today in contrast to Jovian planets, which emit more heat than they receive from the Sun. Third, the surface temperature of a planet with an atmosphere is always greater than the effective temperature - atmospheres that absorb radiation act to warm a planetary surface. Finally, the equations apply to global averages, and not to local variations in the radiative energy budget.

Equations (7.1) and (7.2) are provided because they nicely illustrate four main ways in which planetary climate can be altered. First, the amount of radiation from the star ($S$) can change. The solar constant at Earth varies by only ~0.1% over the course of a solar cycle. But studies of Sun-type stars suggest that the Sun is ~30% brighter today than it was when the terrestrial planet atmospheres first formed (Fig. 7.2a; Sagan and Mullen, 1972; see also Ch. 2 in Vol. III). This makes the stability of Earth's climate all the more remarkable, because the amount of energy encountering the top of the atmosphere has changed considerably over time.

Second, changes in the albedo ($A$) of a planet will change the amount of incident energy absorbed by the surface (and atmosphere). Variation in cloud cover, the

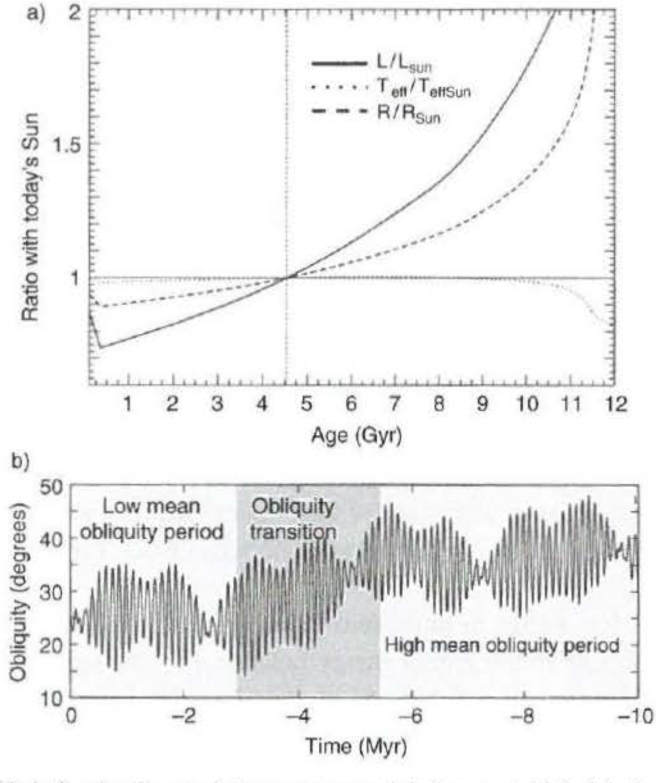


Fig. 7.2 Variation in climate drivers at terrestrial planets. (a) Modeled solar properties (luminosity, effective temperature, radius) as a function of time, relative to today's Sun (Ribas *et al.*, 2005); (b) Martian obliquity (i.e., tilt) as a function of time. (From Levrard *et al.*, 2004.)

extent of polar ices, vegetation, or wind blown dust, for example, can all change the albedos of the terrestrial planets, and will have an influence on the atmospheric energy budget. Venus has an albedo of $\sim$0.9, while the albedos of Earth ($\sim$0.3) and Mars ($\sim$0.25) are considerably lower. Thus Earth and Mars absorb a larger fraction of incident sunlight than Venus. There are of course significant local variations in albedo and in the corresponding absorbed energy at different regions on the surface – especially at Earth.

Third, characteristics of a planet's orbit and rotation influence its energy budget. The amount of solar radiation encountering a planet varies with average orbital distance ($d$), with the result that Venus encounters roughly double the energy that Earth does, while Mars encounters $\sim$45%. Ellipticity of the orbit (not captured explicitly in Eqs. (7.1) or (7.2)) influences variations in incident energy over a given orbit. For example, while the average energy incident at the top of the Martian atmosphere is $\sim$45% that of Earth, it varies between 36% and 52% over a Martian year due to Mars' relatively high orbital ellipticity. This explains why the southern summer at Mars (near perihelion) is more extreme than northern summer. Tilt also influences the amount of sunlight that reaches each part of a planet's surface, making some portions of the planet cold and other portions warm. This effect influences where ices form at the surface, removing some gases from the atmosphere and changing albedo in some locations. Chaotic changes in the eccentricity, obliquity, and spin precession of Mars (Fig. 7.2b) and Earth over periods of tens to hundreds of thousands of years are thought to contribute to climate variations (see Ch. 11 in Vol. III), though the range of variation in both orbital properties (especially tilt) and climate is estimated to be larger at Mars due to the lack of a large moon (e.g., Laskar *et al.*, 2004).

Fourth, the amount of radiation-absorbing atmosphere (i.e., greenhouse gases) influences surface temperatures. The importance of the atmosphere is evident when we compare the amount of energy actually absorbed by Venus, Earth, and Mars. Considering the solar irradiance, average orbital distance, and globally averaged albedo, Earth absorbs the most incident sunlight of the terrestrial planets ($\sim$410 W/m$^2$), followed by a highly reflective Venus ($\sim$270 W/m$^2$), and then the more-distant Mars ($\sim$150 W/m$^2$). It may seem surprising at first that the absorbed energy at Earth's surface exceeds that of the much hotter Venus. However, a planet not only absorbs (and reflects) incident energy, it also radiates energy away. Greenhouse gases such as $CO_2$ and $H_2O$ are efficient at absorbing the infrared energy emitted by the planet, keeping the lower atmosphere warmer than it would be in their absence. The more greenhouse gases are present, the more the surface is warmed, as shown in Eq. (7.2). The thick $CO_2$ atmosphere of Venus provides more than 500 K of greenhouse warming compared to the theoretical surface temperature in the absence of an atmosphere. Earth's atmosphere provides approximately 30 K of greenhouse warming. This warming, while much smaller than at Venus, is crucial to keeping our average surface temperature above the freezing point of water, making life and many aspects of our climate possible. The atmosphere of Mars, while dominated by $CO_2$, is too thin to provide substantial greenhouse warming today. The temperature is warmed only $\sim$5 K due to greenhouse gases. When we take into account the amount of greenhouse gases present in the atmospheres, we see that Venus retains a larger fraction of its radiated heat than the other planets, keeping its surface warmer despite its high albedo. Note also that the high albedo of Venus is due to its extensive cloud cover, which is a product of its climate. Thus albedo and greenhouse gas abundance are linked.

With this context in mind, we can examine how heliophysical processes may contribute to climate variation. They certainly should not influence the orbital characteristics of a planet in our solar system, though one can imagine the strength and timing of a stellar wind influencing planetary migration, and thus orbital distances, in other systems. They may indirectly influence albedo, by altering atmospheric chemistry and promoting the formation of clouds. Certainly, stellar properties directly influence the radiation from a star. And, perhaps surprisingly, interactions between a planet and its host star may change the abundance of climatically important atmospheric gases. This last idea involves an especially rich array of heliophysical processes, and is a focus for the rest of this chapter.

## 7.4 Atmospheric source and loss processes

As described in Sect. 7.2, surface temperature and climate are strongly affected by the amount of greenhouse gases in an atmosphere, which can be viewed as a combination of the total number of particles in an atmosphere (surface pressure) and its composition. A number of mechanisms are capable of changing atmospheric abundance and composition (Fig. 7.3; e.g., Hunten, 1992), only a few of which are heliophysical. We briefly describe them here.

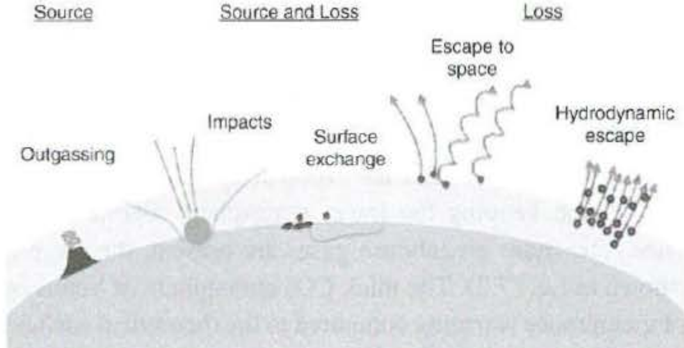


Fig. 7.3 Source and loss mechanisms for planetary atmospheres.

Volcanic outgassing from planetary interiors is thought to be the primary source for the terrestrial planet atmospheres we observe today. Water vapor is the most common gas released in terrestrial eruptions, followed by $CO_2$. Other commonly released gases include sulfur dioxide, nitrogen, argon, methane, and hydrogen. Outgassing should be a declining source of atmospheric particles over solar-system history, as the interior heat required to generate volcanic activity declines. Earth, as the largest terrestrial planet and therefore the one with the most interior heat, has the most evidence for volcanic activity today. Changes in both high-altitude atmospheric sulfur content and thermal emission from the surface suggest that Venus may still be active as well (Smrekar *et al.*, 2010; Marcq *et al.*, 2013). There is no direct evidence that Mars is active today, though relatively young lava flows suggest it could have been active as recently as a few tens of millions of years ago (Hauber *et al.*, 2011).

Atoms and molecules can be exchanged between a planet's surface layers and its atmosphere via a variety of processes and over many time scales. For example, changes in temperature can increase condensation rates to the surface, forming surface liquids or ices (evident on Earth and Mars). Chemical reactions (weathering) can also remove particles from the atmosphere, and is typically most effective in warm or wet environments (evident on Venus, Earth, and Mars). Adsorption removes atmospheric particles that stick to surface materials. Most or all of these processes can be considered to be reversible. Release of particles back to the atmosphere can involve changes in temperature, chemical reactions (including reactions with sunlight), and geologic events that allow subsurface reservoirs access to the atmosphere.

All planetary atmospheres are subject to impact from asteroids, comets, dust, and even atoms and molecules. Impactors of all sizes can deliver volatile species to an atmosphere (e.g., impact delivery is responsible for at least part of Earth's water inventory as well as meteoritic layers observed in terrestrial planet ionospheres). Impacts can also remove atmospheric particles via collisions, and sufficiently large impactors can additionally accelerate atmospheric particles via impact vapor plumes and lofted surface material (ejecta). Because the details of each impact determine whether there is a net gain or loss of atmospheric particles, it is not entirely clear how impacts have contributed to changes in atmospheric abundance and composition over time. It is certain that the importance of impacts has declined with time as the impactor flux decreases. Monte Carlo simulations suggest impacts have resulted in a net gain of atmospheric gases for Earth and Mars over solar-system history, and a net loss for Venus (Heath and Brain, 2014).

Hydrodynamic escape occurs when a light species escapes (thermally – see Sect. 7.6) in sufficient abundance that it becomes equivalent to a net upward wind, and drags heavier species with it through collisions. This process is usually enabled by high solar EUV flux or another form of heating. It should have been significant for all of the terrestrial planets during the first few hundred million years after formation, stripping away most of their primordial atmospheres. The present atmospheres, then, did not form in place on the terrestrial planets; instead they are the product of outgassing and impact delivery, along with subsequent evolution. Hydrodynamic escape may be ongoing at several observed exoplanets today (e.g., Vidal-Madjar *et al.*, 2003; Tian *et al.*, 2005), and perhaps even a few objects in our outer solar system (Tian and Toon, 2005; Strobel, 2009).

The removal of atmospheric particles to space from the upper layers of the atmosphere is commonly referred to as escape to space. This term typically excludes impacts by asteroids, meteoroids, and comets, and hydrodynamic escape is also often listed as a distinct process. Here, escape to space encompasses a set of approximately six processes, all of which provide escape energy to atmospheric particles. The energy is ultimately provided (sometimes directly, and sometimes indirectly) through interaction with the parent star and stellar wind. Escape to space is the most directly related to contemporary heliophysical processes of any of the atmospheric source and loss mechanisms, and is described in more detail in Sect. 7.6. It is currently thought that atmospheric escape has played an important role in the evolution of the climates of both Venus and Mars by altering atmospheric pressure and trace gas abundance.

### 7.5 Requirements and reservoirs for atmospheric escape to space

The removal of atmospheric particles via interactions with the Sun and solar wind is a scientific topic of much debate at present. There is little question that the processes occur; areas of investigation instead focus on whether escape to space is important for planetary evolution. Here, we describe the requirements for escape and reservoirs for escape in planetary atmospheres, and follow in Sect. 7.6 with a discussion of the mechanisms for removing particles.

All particles escaping from a planetary atmosphere share three characteristics. The first is that they have sufficient energy to escape the gravity of the planet. One can easily compute the necessary energy by assuming that a particle can escape when its kinetic energy exceeds its gravitational potential energy. Thus

$$v_{\rm esc} = \sqrt{2GM/r}, \tag{7.3}$$

where $v_{\rm esc}$ is the escape speed, $G$ is the universal gravitational constant, $M$ is the mass of the planet, and $r$ is the radial distance from which the escape occurs (typically the exobase, discussed below). Note that all terrestrial planets have similar bulk densities, so that we can replace $M$ with density times volume, $\rho V$. Volume is proportional to $r^3$, so that $v_{\rm esc}$ scales with radius. Table 7.2 shows typical escape

Table 7.2 *Escape velocities $v_{esc}$ and escape energies E for protons and atomic oxygen*

| | Venus | Earth | Mars |
|---|---|---|---|
| $v_{esc}$ | 10 km/s | 11 km/s | 5 km/s |
| $E(H^+)$ | 0.5 eV | 0.6 eV | 0.1 eV |
| E(O) | 9 eV | 10 eV | 2 eV |

speeds for Venus, Earth, and Mars. Mars has a much lower escape speed because the planet is smaller than Earth or Venus.

Because escape speed is independent of a particle's mass, escape energy must therefore be mass dependent. The table shows typical escape speeds for protons and atomic oxygen (both frequently considered in studies of escape). Escape energies range from fractions of an eV to ~10 eV. Less-massive species require less energy to be removed from an atmosphere.

A second characteristic of an escaping particle is that it is unlikely to collide with other particles after acquiring sufficient escape energy. In planetary atmospheres, the region above which collisions are unlikely is termed the exobase, and is loosely defined as the location where the mean free path of a particle is equal to an atmospheric scale height

$$\frac{1}{n\sigma} = \frac{kT}{mg}, \tag{7.4}$$

where $n$ is the number density of atmospheric particles, $\sigma$ is the cross section for collisions, $k$ is the Boltzmann constant, $T$ is the atmospheric temperature, $m$ is the average mass of atmospheric particles, and $g$ is the local gravitational acceleration. An atmospheric particle must also not be directed downward, so that one can assume that approximately half of all particles given escape energy at or above the exobase location will eventually escape. In reality the exobase is not a sharp boundary; collisions still occur above the exobase, and particles below the exobase can still escape.

Finally, any escaping particles must not be confined to the planet by planetary magnetic fields. This requires either that an escaping particle be neutral, that the planet lack a magnetic field, or that any magnetic fields are weak enough that energized charged particles are able to easily traverse magnetic field lines. Venus lacks a measurable global magnetic field like that of Earth. Mars also lacks a global magnetic field but possesses localized regions of strongly magnetized crust that may locally trap energized atmospheric ions.

Owing to the highly collisional nature of planetary lower atmospheres, escape is generally limited to three regions of the upper atmosphere: the thermosphere,

Table 7.3 *Vertical extent and important species for upper atmospheric regions of terrestrial planets*

| | Venus | Earth | Mars |
|---|---|---|---|
| Thermosphere | ~120–250 km | ~85–500 km | ~80–200 km |
| | $CO_2$, CO, O, $N_2$ | $O_2$, He, $N_2$ | $CO_2$, $N_2$, CO |
| Ionosphere | ~150–300 km | ~75–1000 km | ~80–450 km |
| | $O_2^+$, $O^+$, $H^+$ | $NO^+$, $O^+$, $H^+$ | $O_2^+$, $O^+$, $H^+$ |
| Exosphere | ~250–8000 km | ~500–10 000 km | ~200–30 000 km |
| | H | H, (He, $CO_2$, O) | H, (O) |

the exosphere, and the ionosphere. The altitude and composition of these regions are summarized for each planet in Table 7.3, and the regions are described more generically below.

The thermosphere is a region of neutral particles extending from the mesopause up to the exobase, with temperature that increases as a function of altitude due to X-ray and EUV input from the Sun. The homopause separates the lower, well-mixed thermosphere from a large region where each species is in a separate diffusive equilibrium. Above the homopause, diffusive mixing is slower than gravitational separation, so each gas will take on its own independent scale height based on its mass. Thus, density falls off less quickly with altitude for less massive species (and one can now begin to see why hydrogen was preferentially removed from the top of Venus' atmosphere compared to deuterium, as discussed in Sect. 7.2). The thermosphere is a collisional region, and escape from deep within the thermosphere is unlikely. However, any neutral particle that reaches the top of thermosphere with escape energy should be removed from the atmosphere if it has an outward trajectory.

Thermospheric particles reaching an altitude where collisions are rare and having energy less than the escape energy will populate an extended exosphere or corona on ballistic trajectories. The exosphere is a region of neutral particles extending from the exobase upward to altitudes as high as tens of thousands of km. It is a non-collisional region (by definition), and so any upward-directed particle within it that has velocity greater than the escape velocity is very likely to be removed from the atmosphere. Particles with insufficient energy to escape will return to the exobase on ballistic trajectories unless they are first ionized.

The ionosphere is a region of ionized atmospheric particles extending from the exosphere down into the lower atmosphere (Ch. 13 in Vol. III). Ionospheres have a main density peak produced primarily by photoionization of thermospheric and exospheric neutrals by solar EUV. Other sources of ionization include photoionization by solar X-rays, impact by precipitating particles, and charge exchange with precipitating ions. Unlike neutrals, ionospheric particles can be accelerated by magnetic and electric fields. For planets lacking global magnetic fields, some ions reaching the exobase from below may be accelerated away from the planet by the passing solar wind, even if they initially lack escape energy. Others may re-impact the collisional atmosphere at the exobase.

## 7.6 Atmospheric escape processes and rates

A number of mechanisms are capable of giving atmospheric particles sufficient energy to escape from a planet. They are illustrated schematically in Fig. 7.4 and described below.

Neutral particles can escape an atmosphere in one of three ways: Jeans escape, photochemical escape, and sputtering. Jeans (or thermal) escape occurs because some fraction of neutral particles near the exobase will have sufficient energy to escape simply because the particles have a thermal distribution. Neutral temperatures near the exobase of all three planets are sufficiently low ($\sim$250–1000 K) that only species with small mass (H, D, and He) can escape via this mechanism in significant quantity. The process should be more efficient for Mars (due to its low gravity) and for Earth (due to its higher exobase temperature) than for Venus.

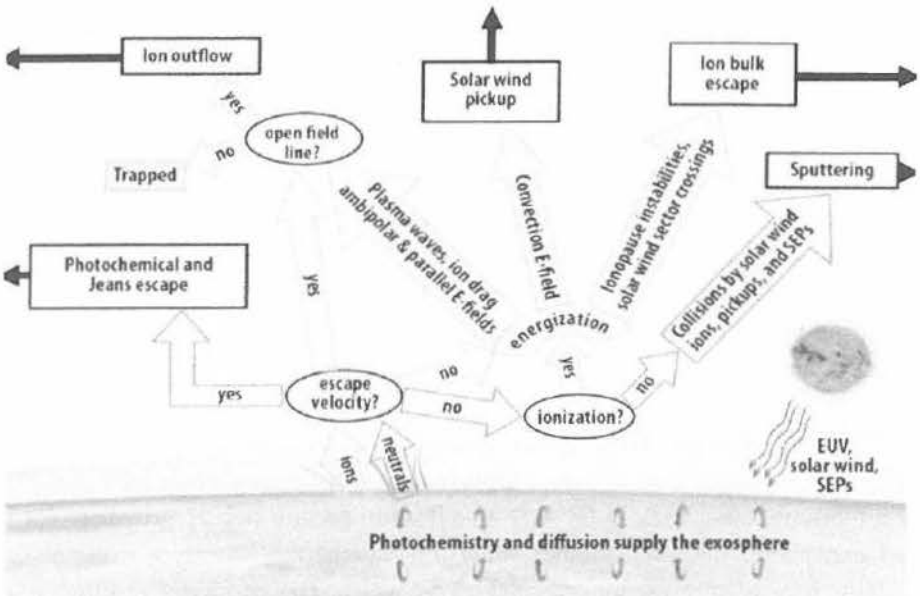


Fig. 7.4 Flowchart showing pathways to energization and escape of particles from a planetary atmosphere. Though Mars is depicted in the diagram, the six processes shown in the rectangular boxes are generic, and can apply to any planet (courtesy NASA MAVEN).

Photochemical escape refers to the escape of fast neutral particles energized by sunlight-driven chemical reactions. These reactions typically involve dissociative recombination of an ionized molecule with a nearby electron, resulting in two fast neutral atoms. Photochemical escape fluxes depend upon ionospheric molecular densities near the exobase, as well as electron density and temperature. Photochemistry is thought to be the dominant loss process for neutral species more massive than hydrogen and helium at Mars. Fast atoms produced photochemically at Venus and Earth are typically not energetic enough to escape the larger gravity.

Atmospheric sputtering occurs when atmospheric particles near the exobase receive sufficient energy from collisions to escape. Collisions occur when energetic incident particles (often ionospheric particles accelerated by electric fields near the planet) encounter the exobase. There are no unambiguous observations that sputtering is actively occurring at any of the terrestrial planets, or contributes significantly to the present-day atmospheric escape rate. However, estimates of loss rates due to sputtering for more extreme conditions suggest it may have been important earlier in solar-system history, especially for unmagnetized planets (Johnson, 1994).

Escaping ions have been directly measured at all three planets, and simulations of the solar-wind interaction with the plasma environments of all three planets are capable of predicting ion escape rates under various input conditions. A number of processes have been identified by which ions can escape an atmosphere, and different authors classify these processes in different ways. Here, we classify ion-loss processes into three categories: ion outflow, ion pickup, and bulk plasma escape.

Ion outflow refers to the acceleration of low-energy particles out of the ionosphere via plasma heating and outward directed charge separation (ambipolar) electric fields. In this case the ion acceleration can occur below the exobase, where collisions maintain a more fluid-like behavior. Ion outflow is the only significant ion loss process for the terrestrial atmosphere, and encompasses a number of processes referred to in the terrestrial literature, including wave heating, polar wind, and auroral outflow (Moore and Khazanov, 2010). Many of these processes should have analogs in the ionospheres of Venus and Mars, and in the localized crustal magnetic fields at Mars.

Ion pickup refers to the situation where a neutral particle is ionized (via photons, electron impact, or charge exchange) and accelerated away from the planet by a motional electric field ($\mathbf{E} = -\mathbf{v} \times \mathbf{B}$). Ion pickup occurs primarily for ionized exospheric neutrals (though some ionized thermospheric neutrals near the exobase region may escape via pickup as well). The motional electric field is usually supplied by the solar wind, so that the process is most relevant for compact magnetospheres unshielded by strong planetary magnetic fields (Venus and Mars) where the solar wind has access to exospheric regions with non-negligible density.

Bulk escape refers to any process which removes spatially localized regions of the ionosphere en masse. Bulk escape is relevant for unmagnetized planets, where the external plasma flow can create magnetic and/or velocity shear with the ionosphere. A popular example involves the Kelvin–Helmholtz (K-H) instability, which may form at the ionopause of Venus or Mars and steepen into waves that eventually detach from the ionosphere (Elphic and Ershkovich, 1984; Penz *et al.*, 2004). Other bulk escape processes are possible as well, such as transport via plasmoid-style flux ropes that may remove ionospheric plasma from Martian crustal magnetic-field regions (Brain *et al.*, 2010a).

It is convenient to think of ion-escape processes in terms of the type of electric field responsible for their removal. A simplified version of Ohm's Law describes the most important electric field terms that influence ion motion

$$\mathbf{E} = -(\mathbf{v} \times \mathbf{B}) + \frac{1}{ne}\mathbf{J} \times \mathbf{B} - \frac{1}{ne}\nabla \mathbf{P}_e, \tag{7.5}$$

where $\mathbf{E}$ is the total electric field, $\mathbf{v}$ is the plasma bulk velocity, $\mathbf{B}$ is the magnetic field, $\mathbf{J}$ is the current density, $\mathbf{P}_e$ is the electron pressure tensor, $n$ is the plasma number density, and $e$ is the electron charge. The three terms on the right-hand side of the equation are the motional electric field, the Hall electric field, and the electron pressure gradient. The Ohmic and electron inertial terms have been neglected. There is varying overlap of these three terms with the ion-escape processes described above. This alternate classification scheme has the advantage that each term can be evaluated unambiguously in global simulations, and compared with observations. It also highlights that a combination of processes can be responsible for removing ions from planetary atmospheres.

A number of estimates of escape rates have been produced for all three terrestrial planets, based both on observations and simulations. In broad terms, the present day global escape rate for Venus is estimated to be $10^{24}$–$10^{26}\,\mathrm{s}^{-1}$ (Lammer *et al.*, 2009). The escape rate for Earth is $10^{25}$–$10^{27}\,\mathrm{s}^{-1}$, and for Mars is $10^{24}$–$10^{26}\,\mathrm{s}^{-1}$.

Two aspects of these estimates should strike the reader. First, the numbers appear to be very large. However, when we consider that the surface areas of the terrestrial planets are on the order of $10^{18}\,\mathrm{cm}^2$, we see that the escape rates are on the order of $10^6$–$10^9\,\mathrm{cm}^{-2}\,\mathrm{s}^{-1}$. In contrast, atmospheric densities near the surface range from $10^{17}$–$10^{20}\,\mathrm{cm}^{-3}$, and column densities range from $10^{23}$–$10^{27}\,\mathrm{cm}^{-2}$. So escape rates are a very small fraction of the number of particles in the present day atmospheres, though accumulated over $\sim$4 billion years ($\sim 10^{17}$ s) they may be substantial. For this latter point the two orders of magnitude uncertainty in escape rates are crucial; they are the difference between heliophysical drivers being the main loss mechanism for planetary atmospheres or merely an afterthought in determining present-day atmospheric abundances.

Second, the escape rates for Venus, Earth, and Mars are all similar, within the admittedly large uncertainties. Given their similarity, we are forced to question a number of common assumptions. Does planetary size or distance from the Sun play a significant role in the removal of atmospheric particles to space? Does the presence of a global magnetic field significantly inhibit escape? And are differences in the present-day escape rates (long after the formation of secondary atmospheres) at all indicative of the amount of escape that has occurred at each body over solar system history? Each of the above questions is ripe for investigation.

Finally, it is important to keep in mind that escape to space not only influences atmospheric abundance but also atmospheric composition, which can be important in planetary evolution. One example is the aridity of the Venus atmosphere. The loss of atmospheric water is attributed to dissociation of the water in the atmosphere by sunlight, and the subsequent escape to space of oxygen. Water is only a trace gas in planetary atmospheres, but is an important greenhouse gas and is extremely important for habitability. So even if escape to space does not appreciably change atmospheric thickness, it may contribute in important ways to climate. Interestingly, the escape rates listed above, when converted to precipitable microns of water, amount to global layers of water only centimeters thick. More than this is assumed to have been lost from Venus, suggesting either that escape rates have changed over time (and are low today) or that other processes (such as impacts) have been important for removing water.

## 7.7 External drivers of escape

Observations, simulations, and common sense all tell us that atmospheric escape rates are not constant, and are influenced by a number of heliophysical drivers that vary on both short and long time scales. The reader is also referred to Chs. 2 and 11 in Vol. III.

The three main drivers are photons, charged particles, and electromagnetic fields. Photons deposit energy in atmospheres when they are absorbed by atmospheric particles. Extreme UltraViolet (EUV) and soft X-ray wavelengths (generated in the solar corona and chromosphere, and not to be confused with solar luminosity) provide the dominant energy source in upper-atmospheric regions. Charged particles in the solar wind also supply energy to planetary upper atmospheres and plasma environments. Table 7.4 summarizes some of the relevant quantities of the solar wind at each terrestrial planet. While density and velocity can each vary independently, studies of solar wind influences on atmospheric escape (especially the induced magnetospheres of Venus and Mars) typically use solar wind pressure ($\rho v^2$) as the organizing quantity. Finally, the solar wind carries a magnetic field, which creates a convection electric field ($\mathbf{E}_{SW}$) in the frame of the planet that

Table 7.4 *Typical properties of the solar wind (SW) and interplanetary magnetic field (IMF) at terrestrial planets*

| | Venus | Earth | Mars |
|---|---|---|---|
| IMF strength | 10–12 nT | 6 nT | 3 nT |
| Solar-wind speed | 400 km/s | 400 km/s | 400 km/s |
| Solar-wind density | 10–15 $cm^{-3}$ | 6 $cm^{-3}$ | 1–3 $cm^{-3}$ |
| Alfvén speed | 70 km/s | 55 km/s | 45 km/s |
| Mach number | 5–7 | 6–8 | 8–10 |
| SW $H^+$ gyroradius | 1500 km | 2500 km | 5000 km |
| SW $H^+$ gyroradius/$R_P$ | 0.5 | 0.4 | 3 |

depends upon solar wind velocity and interplanetary magnetic field (IMF) strength and orientation (see Eq. (7.5)). Magnetic and electric fields organize charged particle motion, and electric fields accelerate charged particles; both effects influence the ability of charged particles to escape a planet's atmosphere.

The external drivers of atmospheric escape vary on four main time scales. Billion year time scales are associated with the age of the Sun, and both theoretical calculations and observations of Sun-type stars suggest that all three drivers should have declined in intensity with age (Fig. 7.5; see Chs. 2 and 11 in Vol. III). EUV flux varies by factors of several over a solar cycle (from solar minimum to solar maximum), and solar wind pressure varies by factors of 2–10. The IMF, in particular, varies with the solar rotation period, and all three drivers also vary on more rapid time scales of minutes to hours.

Variability in the heliophysical drivers should influence atmospheric escape rates. In general, an increase in solar EUV fluxes (e.g., a transition from solar minimum to solar maximum) is expected to result in an increase in loss rates of neutral particles. Energy from solar photons heats the upper atmospheric neutrals, so that Jeans escape rates should increase with solar EUV. This is likely to be true at Mars, but not at Earth where hydrogen escape from the exobase is limited not by the available energy, but by the supply (via diffusion) of particles from lower altitudes (see discussion in Tian *et al.*, 2013). Jeans escape should be negligible at Venus today, but may have been significant in the past if either exobase temperatures or solar EUV fluxes were much higher. Energy from solar photons is also used to drive the chemical reactions necessary for photochemical escape, so that contemporary Martian photochemical escape should vary with EUV flux. Neutral escape rates should be largely insensitive to changes in both the solar wind and the IMF, except for sputtering rates from Venus and Mars, which are thought to be dominated by re-impacting atmospheric pickup ions and will therefore increase as the pickup-ion population increases in response to changes in solar EUV.

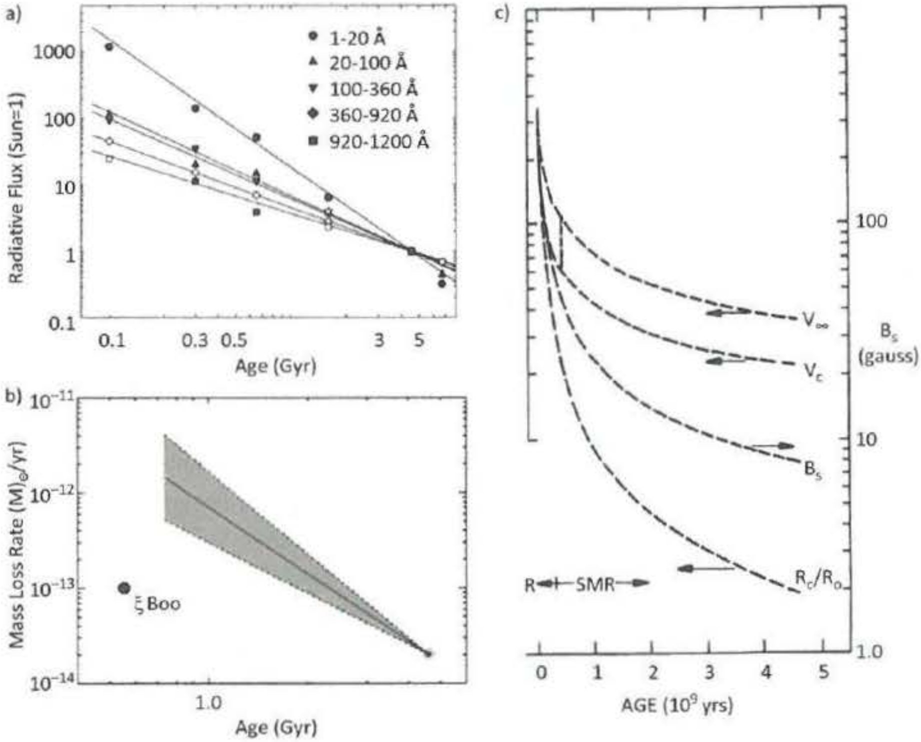


Fig. 7.5 Evolution of solar drivers of atmospheric escape. (a) Solar EUV photon flux, relative to today (from Ribas *et al.*, 2005); (b) solar mass loss rate (i.e., solar wind flux) (from Wood *et al.*, 2005); (c) interplanetary magnetic field (curve labeled $B_S$; from Newkirk, 1980).

Ion escape rates should also vary with the three drivers. An increase in solar-wind pressure will cause a corresponding decrease in the size of the magnetospheric cavity at all terrestrial planets, effectively lowering the pressure balance altitude between the solar wind and planetary obstacle to the flow. For Mars, with an extended neutral corona, an increase in solar-wind pressure exposes significant additional high-altitude neutrals to ionization and stripping by the solar wind (via electron impact and charge exchange). The IMF, by contrast, chiefly organizes the trajectories of escaping particles at Venus and Mars; large gyroradius pickup ions are preferentially accelerated away from the planet in regions where $\mathbf{E}_{SW}$ points away from the planet. At Earth, the orientation of the IMF affects the location and extent of cusp regions, from which outflowing ions escape. EUV fluxes have a more indirect effect. In total, one might expect the ion escape rate to increase at solar maximum due to the additional energy input from EUV. At unmagnetized Venus and Mars, however, the increased ionospheric content deflects the solar wind around the planet at higher altitudes and can prevent the interplanetary magnetic

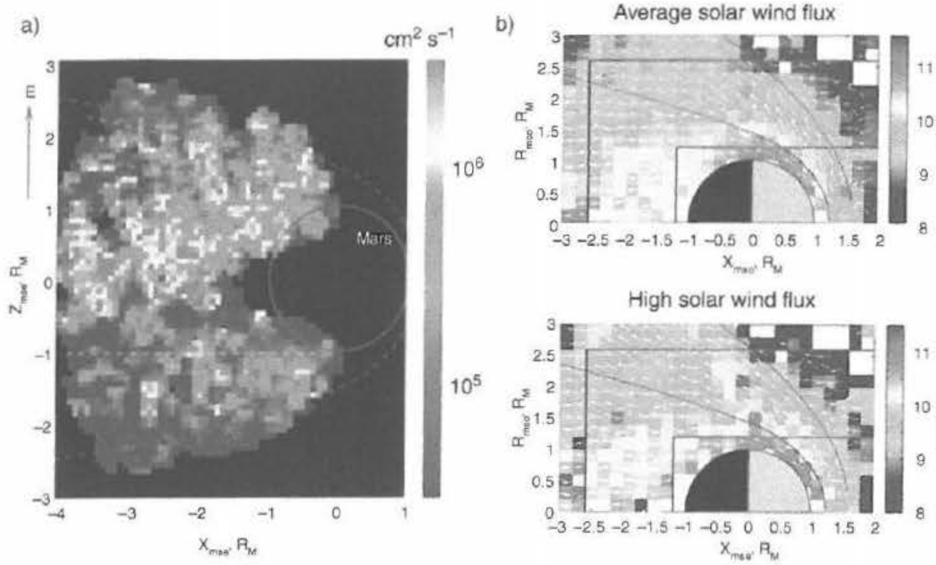


Fig. 7.6 Ion escape from the Martian atmosphere, organized by solar drivers. The Sun is to the right in both panels. (a) Escaping ion fluxes downstream from Mars are greater in the hemisphere of upward directed (with respect to the planet) solar wind electric field (Barabash *et al.*, 2007); (b) escaping ion fluxes downstream from Mars are greater during periods of high solar wind flux. (From Nilsson *et al.*, 2011.) A black and white version of this figure will appear in some formats. For the color version, please refer to the plate section.

field from entering the ionosphere. The escape of heavy ion species (which are concentrated at lower ionospheric altitudes) via pickup and bulk escape may therefore remain roughly constant, or even decrease during solar maximum periods, even as lighter ion species escape more efficiently.

Observations support the above assertions in a general sense, though quantification of many of the effects is still being teased out of the available data. Here, we take ion escape from Mars as an example. First, the organization of escaping ions by the IMF is borne out by observations (Fig. 7.6a). Next, the flux of escaping planetary ions has been correlated with the solar wind intensity (Fig. 7.6b). Finally, the fluxes of escaping ions measured at solar minimum and solar maximum differ by approximately an order of magnitude (Lundin *et al.*, 1990; Barabash *et al.*, 2007). These results should be cautiously interpreted, however, because the measurements were made by two spacecraft with different orbits and instruments.

Models also support these trends, and provide a useful complement to spacecraft observations because they are able to perform controlled experiments on how a planet responds when only one external driver is changed. Further, models are not limited by a spacecraft orbit or instrument observing geometry; they provide

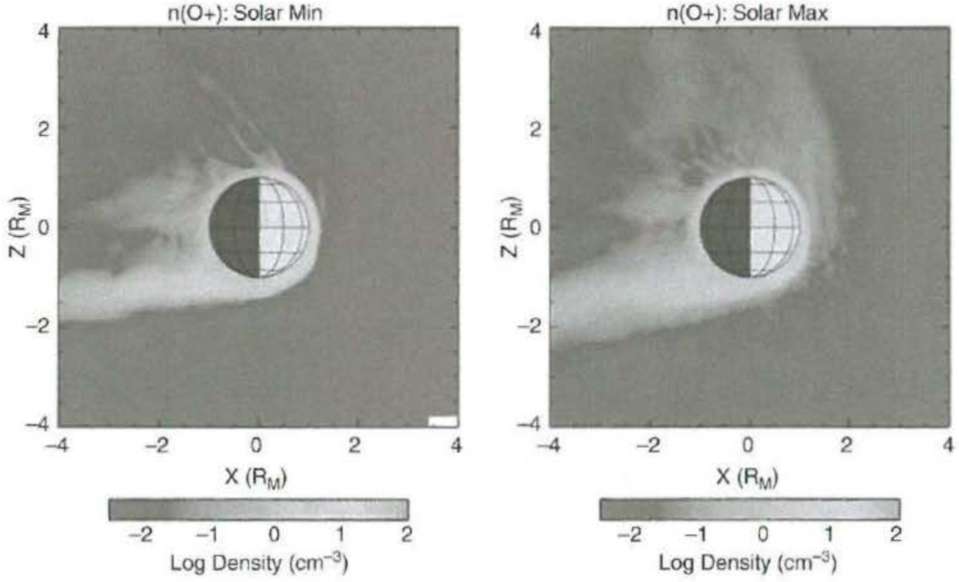


Fig. 7.7 Density of escaping atomic oxygen ions from Mars at solar minimum (left) and solar maximum (right) as predicted by a global hybrid plasma simulation. The Sun is to the right in each panel, and the solar wind electric field points toward $+y$. Courtesy E. Kallio and R. Jarvinen.

information about how the entire system responds. Again using ion escape from Mars as an example, a variety of models have been used to simulate the near-space environment and predict atmospheric ion escape rates. The models predict that escape rates increase from solar minimum to solar maximum (Fig. 7.7), and with solar-wind pressure (see discussion in Brain *et al.*, 2015). And the IMF orientation clearly controls the trajectories of escaping ions. The models employ different physical assumptions, boundary conditions, and implementation schemes, and so it is not surprising that the models disagree on the magnitude of each of the above-described effects. A major challenge facing the community at present is determining which models produce results that best match the observations. The answer is likely to depend on location, type of observation, and external conditions.

Solar storm periods provide extreme cases for each of the three drivers mentioned above, and in a sense provide a window into conditions earlier in solar-system history. Solar flares, solar energetic particle (SEP) events, and the enhanced magnetic field associated with coronal mass ejections (CMEs) have all been measured at the three terrestrial planets. Initial efforts at quantifying the effect of solar storm periods suggest that ion escape rates can increase by an order of magnitude or more at Mars and Venus (Futaana *et al.*, 2008), and suggest that Earth's escape rates increase less during similar periods (Wei *et al.*, 2012). However, much more work remains to be done on this topic, especially with regard to comparing the responses of the different planets to solar storms.

## 7.8 Internal drivers of escape

A number of characteristics of a terrestrial planet itself influence the properties and energetics of upper atmospheric reservoirs for escape, including transient events such as dust storms (e.g., for Mars), or longer-lived phenomena such as gravity waves that couple the lower and upper atmospheres. In the context of heliophysics, the nature of a planet's intrinsic magnetic field is of the greatest relevance.

Earth possesses a global dynamo magnetic field today (see Ch. 6 in this volume and Ch. 7 in Vol. III), while Venus lacks a measurable dynamo field. Mars also lacks a dynamo magnetic field but has crustal magnetic fields (Acuña *et al.*, 1998) that have significant influence on the upper atmosphere and plasma environment. The strength of the crustal fields and their higher concentration in the more ancient southern hemisphere suggest that they formed in the presence of an ancient global dynamo magnetic field, which shut off as many as 4.1 billion years ago (Lillis *et al.*, 2008). Temperature gradients in planetary interiors are important for driving the convection necessary to sustain a dynamo; in simple terms the smaller Mars cooled more quickly than Earth and became incapable of supporting an internal dynamo early in the planet's history. Venus may have hosted a dynamo at one time, but the much hotter surface makes it highly unlikely that any remanent crustal magnetism has been preserved (and orbiting spacecraft have not detected any). Thus we are left today with an intrinsic magnetosphere at Earth that deflects the solar wind at large distances from the planet ($\sim 10\ R_E$), and induced magnetosphere at Venus that deflects the solar wind at much closer distances ($\sim 1.3 R_V$), and a similarly sized (with respect to the planet) induced magnetosphere at Mars punctuated by mini-magnetospheres tied to specific regions of the crust and that rotate with the planet.

How do the different magnetic fields influence ion-escape processes and rates? In general terms, planetary magnetic fields influence escape processes in two main ways (Fig. 7.8). First, they add magnetic pressure from the planet that contributes to balancing the pressure in the solar wind, causing the solar wind to be deflected farther from the planet and thus farther from its atmosphere. Second, magnetic fields can alter the topology of magnetic field lines near the planet, reconnecting with the IMF. The presence of intrinsic magnetic field creates closed field lines that shield an atmosphere from the solar wind and trap ions, and open field lines in cusp regions that enable exchange of energy and particles between the atmosphere and solar wind.

It is reasonable to consider that the three ion-escape processes described in Sect. 7.6 should all be influenced by the presence of a planetary magnetic field.

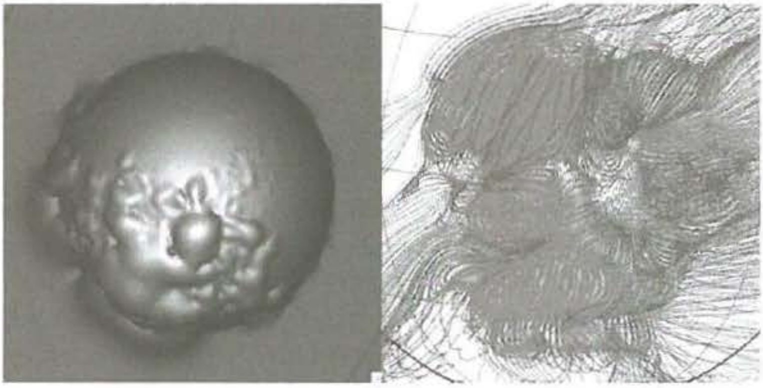

Fig. 7.8 Influence of magnetic fields on planetary near-space environments. Magnetic fields supply magnetic pressure (left: for Martian crustal magnetic fields) that deflect solar wind, but also modify magnetic topology (from Brain, 2006); (right: for the strong Martian crustal fields in the southern hemisphere, where red denotes closed field lines and blue denotes field lines open to the solar wind at one end) that enable exchange of particles and energy between the atmosphere and solar wind. Both renderings result from model calculations that include contributions from crustal fields and external drivers (solar wind or IMF). A black and white version of this figure will appear in some formats. For the color version, please refer to the plate section.

Both ion pickup and bulk plasma escape should be reduced in the presence of a magnetic field, because the solar wind is more effectively shielded from the ionosphere. In addition, ionization of atmospheric neutrals by the solar wind should be reduced in the presence of a magnetic field, decreasing the population of particles available for escape. Ion outflow occurs when cold atmospheric ions are accelerated by electric fields. The electric fields result from a variety of processes, and it is unclear whether outflow should decrease or increase in the presence of a magnetic field. However, intrinsic fields create vertically oriented flux tubes, which should facilitate vertical ionospheric transport. Several outflow acceleration mechanisms also rely on vertical field lines to form or maintain electric fields that accelerate particles, suggesting that outflow is likely to be more effective when a planet is magnetized.

Of the neutral loss processes, sputtering is most likely to be influenced significantly by planetary magnetic fields. Sputtering is primarily caused by planetary ions re-encountering the exobase. Magnetic fields influence both the trajectories of charged particles, and their formation (by preventing the solar wind from accessing high altitude regions of neutral atmosphere). Thus one would expect sputtering loss to be inhibited in the presence of planetary fields. Jeans escape should not be affected by magnetic fields because it primarily involves solar photons and neutral particles. Photochemical escape may be indirectly affected because the fast neutral particles lost via the process are produced when a planetary ion and electron recombine. Like sputtering, the influence of a planetary magnetic field on ion production and motion may influence the loss rate. However, unlike sputtering, most of the processes involved in photochemical loss typically take place deeper in a planet's atmosphere (at or below the exobase) than sputtering, so influences of a magnetic field may be negated by particle collisions.

When considering the total atmospheric loss from a planet, it has often been assumed that the presence of a magnetic field results in lower escape rates. The argument is that intrinsic magnetic fields prevent the solar wind from directly accessing the atmosphere, shielding it from solar wind-related ion loss. Above, we argued that three of six loss processes (ion pickup, bulk plasma escape, and sputtering) should be less efficient in the presence of planetary magnetic fields, and two more should be unaffected or only weakly changed (Jeans escape and photochemical escape). Further, evidence discussed in Sect. 7.2 suggests that Mars lost substantial atmosphere, and that both Venus and Mars lost atmospheric particles to space more efficiently than Earth. Though there are few planets to compare and comparisons are complicated by the lack of controls (e.g., size, distance from the Sun, etc.), it may be telling that Earth possesses both a global magnetic field and a habitable atmosphere.

However, we mentioned in Sect. 7.6 that the measured atmospheric escape rates for Venus, Earth, and Mars are comparable within the current uncertainties. It has recently been proposed that magnetic fields, rather than shielding a planetary atmosphere from stripping by the solar wind, actually collect solar-wind energy and transfer it to the ionosphere along field lines (Strangeway *et al.*, 2010). Global magnetic field lines converge near the cusps, so that the energy is more spatially concentrated than for unmagnetized planets. The escape rate for a given planet may be comparable when it is magnetized, or even greater because planetary magnetic fields extend much farther than the planet's atmosphere, giving it a larger energy collecting cross section in the solar wind. One key difference with magnetized planets is that the concentrated energy in cusp regions is likely to lead to more efficient removal of heavy species.

There are a few caveats. First, there are large error bars at present on planetary atmospheric escape rates at all three terrestrial planets, so that we are still unsure whether the escape rates are similar. Second, not all solar-wind energy collected by a planet need go into removing atmospheric particles. It could instead drive chemistry or upper-atmospheric dynamics and heating. However, measurements from Earth suggest that the solar-wind Poynting flux scales exponentially with the upward directed ion flux in cusp regions (Fig. 7.9). Finally, accelerated ions in Earth's cusps need not escape the planet at all. It is currently uncertain what

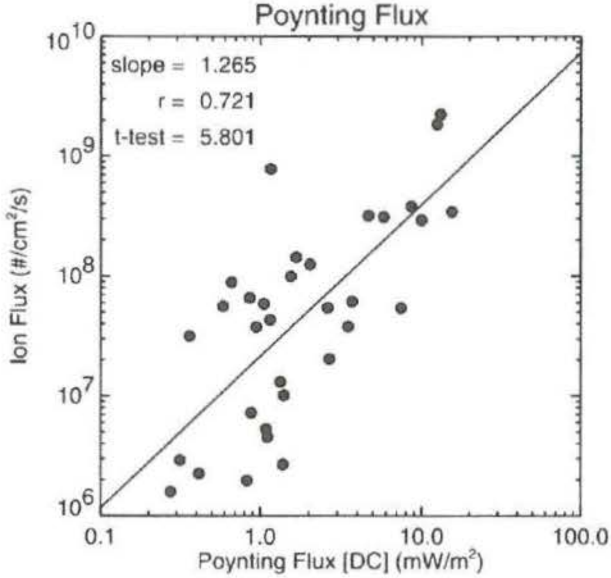


Fig. 7.9 Measurements of outflowing oxygen ions from Earth's cusp regions against solar wind Poynting flux. (From Strangeway *et al.*, 2005.)

fraction of ions leave Earth's magnetosphere, and what fraction is returned to the planet along magnetic field lines. A number of issues must be investigated and resolved before we can determine whether magnetic fields protect an atmosphere from being lost.

## 7.9 Frontiers

It should be apparent that there are multiple likely connections between heliophysics and the climates of terrestrial planets. Yet we are still uncertain whether Mars lost a bar of atmosphere to space, or whether Earth's magnetic field is a shield for the atmosphere. There are clearly exciting and important questions to be tackled in the coming years. Where and how is the community likely to make progress in the next 5–10 years? Below, we discuss three frontier research areas.

First, global simulations of planetary plasma environments have become more and more capable over the past two decades. These models are capable of simulating global magnetospheric interactions with increasingly accurate physical assumptions and spatial resolution as computers become ever faster. Still, very few models are capable of treating a global dipole magnetic field as a knob to be turned on or off in order to study the influence of magnetic fields on atmospheric escape processes. Fluid models for Earth's solar-wind interaction could turn off the planetary dipole, but would not be capable of accurately capturing several of the kinetic ion-escape processes (which may dominate escape from Earth). At least one hybrid (kinetic ions and fluid electrons) model has simulated Mars with and without a global dipole, but the simulated planetary field was made very weak so that the model was computationally tractable, and the model is not optimized for studying escape from low-altitude regions (Kallio and Barabash, 2012). Within the next several years, however, it seems likely that a model will become capable of simulating magnetized and unmagnetized versions of Earth, Venus, and Mars using physical assumptions that allow investigation of all ion-escape processes. Further, current efforts to couple global plasma simulations with models for the exosphere and thermosphere hold promise for capturing all ion and neutral escape processes in a single self-consistent simulation (e.g., Dong *et al.*, 2014).

Next, analyses of existing and ongoing observations should provide useful constraints for models on the importance of individual loss processes at each planet under varying conditions, and the role of magnetic fields. Existing observations of atmospheric escape from Venus, Earth, and Mars are typically presented for a single object. More detailed comparisons between objects are in order. Spacecraft missions such as MAVEN, which arrived at Mars in 2014, will allow investigation of the physical processes that result in escape (Jakosky *et al.*, 2015). MAVEN measures the drivers of escape (solar photons, particles, and fields), the atmospheric reservoirs for escape, and the escaping particles. Earlier missions, while very productive, have been hampered by incomplete observations. Further, MAVEN measures atmospheric escape processes from both magnetized and unmagnetized regions of Mars, which holds promise for comparisons between atmospheric regions that differ only in their magnetic field. These results are likely to be useful in assessing whether magnetic fields reduce escape rates.

Finally, the lessons from models and observations of solar-system objects can be applied to exoplanets. Since 1995, the number of known exoplanets in our Galaxy has grown, from zero into the thousands, as detection methods have improved substantially and dedicated spacecraft missions and telescopes have been commissioned. Now, we find ourselves at a point where we infer that most stars in our Galaxy have planets, and that there may be as many as 40–100 billion habitable exoplanets in the Milky Way (Petigura *et al.*, 2013). The word habitable is tricky, though. The studies that calculated 40–100 billion habitable planets assumed that Venus and Mars were habitable. This is a fair assumption because both planets may have been habitable at their surface at some point in solar-system history. However, the assumption is very probably over-generous, considering Venus and

Mars today. Can we use our understanding of heliophysical processes to determine which planets are likely to have climates conducive to surface habitability? Can we apply models tuned to Venus, Earth, and Mars, and validated against spacecraft observations of atmospheric-loss processes, to narrow the list? It will certainly be exciting to try.