https://doi.org/10.1142/9781800613140_0004

Chapter 4

Exploration of the Atmospheres of the Terrestrial Planets

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In this chapter, we describe the atmospheres of the terrestrial planets of our Solar System. After a brief introduction on the concept of "comparative planetology", we introduce different theoretical elements which permit us to better grasp the differences existing within the atmospheres of Mercury, Venus, Mars and Earth.

1. Introduction: Comparative Planetology

Planetary atmospheres are gaseous envelopes surrounding celestial bodies. They differ in terms of temperature, pressure and composition, and therefore offer natural test beds to improve our knowledge on the physical, chemical and dynamical processes that can occur in such systems. Indeed one of the main objectives of comparative planetology is to study all atmospheres to better understand our own atmosphere, to understand its past and future evolution. Processes like the interaction of the atmosphere with the solar radiation or with its magnetosphere, exchange with the surface and with space, the dynamical circulation, or even seasonal cycles occur in all atmospheres, but they are shaped by the different conditions occurring on each celestial body, like its temperature and pressure or its proximity to its Sun. Comparing different planetary atmospheres allow to investigate the origin and history of the Solar System. It is possible to retrace its evolution from the collapsing of the rotating disk of gases around the Sun to the current composition of the different atmospheres. Measuring the abundance of noble gases, and in particular that of Helium or Argon, of hydrogen and carbon, provides precious information to validate (or not) different hypothesis. Comparative planetology can also be seen as the first step toward classifying and understanding the atmospheres of the numerous exoplanets that have been and will be detected.

To illustrate the concept of comparative planetology, let's consider the seasonal effect on Mars, Venus and Earth. Seasons on Mars and Earth are due to the high inclination of their rotation axis from the vertical of the plane of ecliptic. Seasons on Mars are exacerbated by its elongated elliptical orbit around the Sun, whereas that of Earth is almost circular. On Venus no seasons are expected because of its quasi-circular orbit and its very small inclination.

Early Venus radio observations from Earth published in 1958 showed an amazingly hot temperature, upwards of 600 K, which was confirmed by the flybys of *Mariner 2* in 1962. This high temperature could not be explained at that time. Slowly the idea of an exceptional greenhouse effect emerged. Scientists realized that the atmosphere of Venus was filled not only with CO_2 but also with an opaque haze. Its nature was unknown, and in the 1960s scientists could only say that the haze was probably caused by some kind of tiny particles. At that time it was believed that the unraveling of the precise role of aerosols in the Venus atmosphere would certainly benefit studies of chemical contamination of Earth's atmosphere.[1] In the early 1970s, ground-based telescope observations produced extraordinarily precise data on the optical properties of these aerosols, and at last they were identified. The haze was made of sulfur compounds. The greenhouse effect of the sulfates could be calculated, and by the late 1970s, NASA's findings about sulfate aerosols strengthened the belief that these particles could make a serious difference to the Earth's climate as well. Sulfates were emitted by volcanoes and, increasingly, by human industry, so Venus had things to tell about climate change on Earth.

In 1971, the spacecraft *Mariner 9* settled into orbit around Mars and observed a great dust storm shrouding the entire planet. The observers immediately saw that the dust had profoundly altered the Martian climate, warming the planet by tens of degrees. The dust settled after a few months, but its lesson was clear. Haze, clouds and dust could warm an atmosphere. More generally, anyone studying the climate of any planet would have to consider dust very seriously. In particular, it seemed that on Mars the temporary warming had reinforced a pattern of winds that had kept the dust stirred up. It was a striking demonstration that feed-backs in a planet's atmospheric system could flip weather patterns into a drastically different state. That was no longer speculation but an actual event in full view of scientists.

The present atmosphere of Mars is thin, and its surface is too cold to allow liquid water. Still, the presence of ancient channels supports the belief that, in the past, a warmer climate allowed running water, possibly life sustaining. Venus was originally cooler and had a greater abundance of water. Because the planet was slightly closer to the Sun than the Earth, its water never liquefied and remained in the atmosphere to participate in the greenhouse effect. Being in gaseous form, water could reach higher altitudes where ultraviolet radiation would dissociate the molecule into H and O atoms. One clue to the Mars and Venus atmosphere evolutions is the ratio of deuterated water vapor HDO to normal water vapor H_2O. On Mars the D/H ratio has been measured to be, in the lower atmosphere, six times higher than on Earth. On Venus it is even higher, reaching 120 times the Earth value. Because D atoms are twice heavier than H atoms, their escape rate from the top of an atmosphere is much lower. This leads to believe that there has been in the past more H_2O on Mars or Venus than now. The exact determination of these past abundances depends on the ratio of escape rates of D and H, directly linked to the D/H ratio in the upper atmosphere. The total HDO remaining now in the planet is an indication of how much water the planet could have contained in the past.

Mars has the advantage, compared to Earth, that it has preserved traces of its long gone history. Earth surface is indeed continuously remodeled by erosion and plate tectonics. There is almost no trace left of the time when microorganisms colonized it, before life was governed by the carbon chemistry. Mars could tell us this lost history and could reveal how life emerged.

To conclude this long list of how solving scientific questions on other planets might help scientists to better understand their own world, we could cite H. Newell, who, in 1980 reported that the "study of the role of halogens in the atmosphere of Venus... led to the suspicion that chlorine produced in Earth's stratosphere from the exhausts of Space Shuttle launches or from Freon used at the ground in aerosol sprays might dangerously deplete the ozone layer".[1] Maybe investigating the photochemistry of the PAH in the Mars atmosphere or analyzing the link between CO and the Venus circulation, to take only two examples, might one day give some insights to unsolved problems on Earth.

In the following sections, we will provide some general information on planetary atmospheres and how that information can be obtained through observations.

2. Generalities on Planetary Atmospheres

In this section, we will describe different characteristics shared by all planetary atmospheres, which will help understand their specificities, described in the following sections. The atmosphere usually refers to the gaseous envelope around a planet or body. Initially defined for rocky bodies having a solid surface, the definition has been extended to gaseous objects, such as the giant planets or even stars. In the case of the four giant planets of our Solar System (Jupiter, Saturn, Uranus and Neptune), their atmospheres are dominated by hydrogen and helium. The outer gas phase atmosphere blends into liquid once the atmosphere pressure has increased beyond a critical point. As a consequence, there is no clear boundary between the surface and the atmosphere.

Within any atmosphere, temperature, pressure and density change with altitude. The pressure at a given altitude is the weight of air above that altitude per unit area (Fig. 1). The pressure decrease

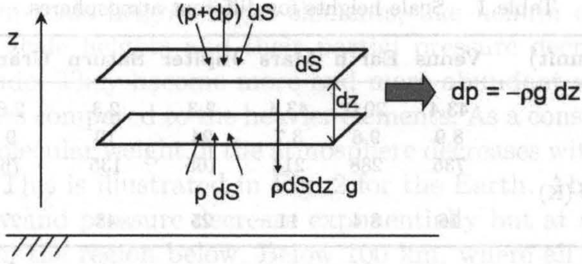

Fig. 1. Definition of atmospheric pressure.

with increasing altitude is described by the hydrostatic law:

$$\frac{dp}{dz} = -\rho(z)g(z), \tag{1}$$

$$p(z) = \int_z^\infty \rho(z)g(z)\,dz, \tag{2}$$

where p is the pressure at altitude z, ρ is the density ($\rho = m/V$, with m the total mass of the species and V the volume occupied by the gas) and g is the acceleration due to gravity. Usually g can be considered constant, since the atmosphere depth is negligible compared to the radius of the planet.

The *equation of state* describes the relationship between pressure, volume and the temperature of a real gas. The ideal gas law is a special case when the gas can be considered as ideal, i.e., as a theoretical gas composed of molecules between which there is no interaction. In this case, the pressure is expressed through the equation:

$$p = \frac{\rho}{m}RT. \tag{3}$$

When combining the hydrostatic law (Eq. (2)) and the ideal gas law (Eq. (3)), the following equation is obtained:

$$\frac{dp}{p} = \frac{g(z)m(z)}{RT(z)}dz = \frac{dz}{H}, \tag{4}$$

with

$$H = \frac{RT(z)}{g(z)m(z)}, \tag{5}$$

where H has the dimensions of a distance, and is called the "scale height". Values of the scale height for different atmospheres are given

Table 1. Scale heights for different atmospheres.

Parameter (unit)	Venus	Earth	Mars	Jupiter	Saturn	Uranus	Neptune
m (g)	43.4	29.0	43.4	2.3	2.3	2.6	2.5
g (m/s^2)	8.9	9.8	3.7	24	10	9	11
Surface temperature (K)	735	288	214	165	135	76	72
H (km)	16	8.4	11	25	48	27	22

in Table 1. They were calculated considering the temperature at the surface of the planet. Scale heights are lower for massive planets and dense atmospheres. The higher the gravity or density or the lower the temperature, the more concentrated the atmosphere is toward the surface. However, the scale height varies with altitude as the temperature also varies. For example, on Earth, the scale height is about 8.4 km at the surface but only 6 km at an altitude of 13 km (210 K).

Considering that the temperature is constant with altitude (or that the scale height is constant), the pressure decreases with altitude according to an exponential law.

$$p(z) = p(0)e^{-z/H}. \tag{6}$$

The scale height is the height above a reference level at which pressure decreases by a factor e (i.e., by about 2.7). The hydrostatic hypothesis states that the vertical acceleration of air masses is zero, not that vertical motion is impossible. This is a reasonable hypothesis on large spatial scales of several kilometers, but not in some kinds of clouds like cumulonimbus, where vertical accelerations can be important.

In most of the atmospheres, all atmospheric constituents are uniformly mixed. Indeed in the lower layers of the atmosphere, convection and turbulent diffusion mix the different gases homogeneously. The hydrostatic law applies to the atmosphere considered as a whole, and all species share the same scale height. This region is called the homosphere. Above a certain altitude, called the homopause, the turbulent diffusion is less important and the different species separated under the effect of gravity according to their mass. The hydrostatic law applies to each species separately and each species has its own scale height. Light elements, like helium or hydrogen, have high scale heights and their partial pressure decreases slowly with altitude. They become more and more abundant with increasing altitudes compared to the heavier elements. As a consequence, the average molecular weight of the atmosphere decreases with increasing altitudes. This is illustrated in Fig. 2 for the Earth. Above 100 km, the density and pressure decrease exponentially but at a rate which differs from the region below. Below 100 km, where all constituents are well mixed, their densities decrease with altitude at the same exponential rate. Above 100 km, the mean free path becomes larger than the turbulent displacement of air. Diffusive transport becomes dominant. Because molecular diffusion depends on the molar weight of the species, species will be stratified differently, heavier molecules decreasing more rapidly with altitude.

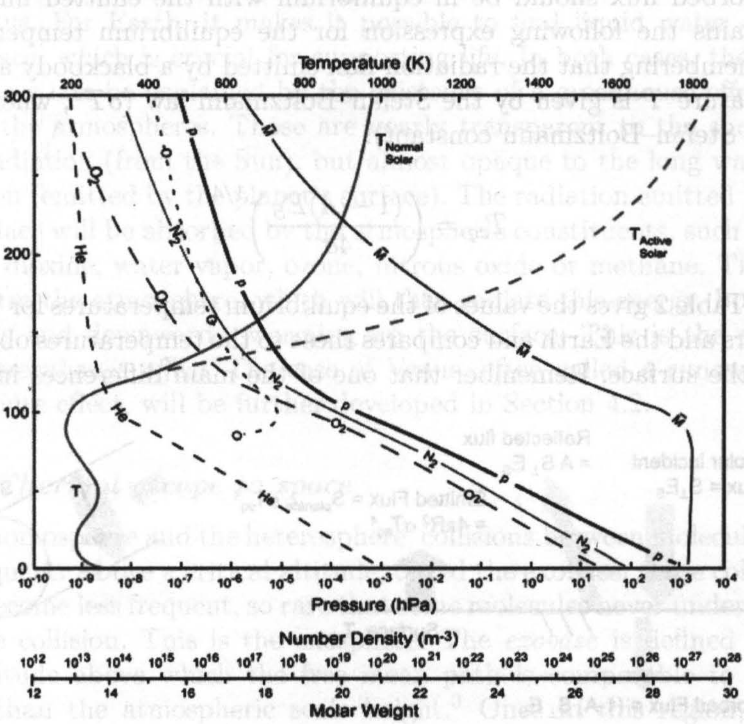

Fig. 2. Global mean pressure (p), temperature (T), mean molar weight ($\overline{M}$) and number densities of several atmospheric constituents of the Earth atmosphere as functions of altitude. Adapted from Ref. 2.

2.1. *Equilibrium temperature and greenhouse effect*

The *equilibrium temperature* of a planet is the theoretical temperature that a planet would be at if it was in radiative equilibrium, i.e., the radiation the planet would emit as a black body at that temperature would compensate the energy received by its parent star. In this simple frame, the planet does not have any atmosphere nor internal heat source.

Let's consider the general case of a planet of radius R, with a surface temperature of T_{eq}, and a planetary albedo A (percentage of the incident radiation which is reflected). Figure 3 represents schematically the different energetic fluxes: the incident solar flux, the fractions of this flux reflected and absorbed by the planet, and the black body flux emitted by the surface. By expressing that the absorbed flux should be in equilibrium with the emitted flux, one obtains the following expression for the equilibrium temperature, remembering that the radiation flux emitted by a blackbody at temperature T is given by the Stefan–Boltzmann law (σT^4, where σ is the Stefan–Boltzmann constant):

$$T_{eq} = \left(\frac{(1-A)E_S}{4\sigma} \right)^{1/4}. \tag{7}$$

Table 2 gives the values of the equilibrium temperatures for Venus, Mars and the Earth and compares these to the temperatures observed at the surface. Remember that one of the main differences in these

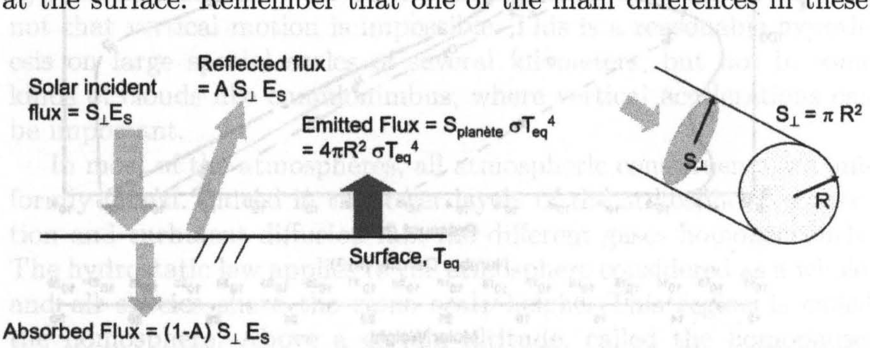

Fig. 3. Equilibrium temperature: (Left) Description of the fluxes; (Right) Definition of the surface intercepting the incoming solar radiation.

Table 2. Comparison between equilibrium and surface temperatures for some planets.

Parameter	**Venus**	**Earth**	**Mars**
Bond albedo	0.75	0.31	0.25
Incident flux (W/m^2)	2621	1370	590
Equilibrium temperature (K)	232	254	210
Mean surface temperature (K)	737	288	215
Excess of temperature (K)	505	34	$\tilde{0}$

temperatures is the existence of an atmosphere surrounding the planets. For Mars, there is almost no difference, which is not surprising given the thinness of the martian atmosphere, as we will see in Section 4. The differences are quite significant for Earth and even more for Venus. For Earth, it makes it possible to find liquid water on its surface, which is crucial for supporting life. In both cases, these differences can be explained by the existence of a greenhouse effect within the atmospheres. Those are nearly transparent to the short wave radiation (from the Sun), but almost opaque to the long wave radiation (emitted by the planet's surface). The radiation emitted by the surface will be absorbed by the atmospheric constituents, such as carbon dioxide, water vapor, ozone, nitrous oxide or methane. This will warm the atmosphere, which will then radiate this energy, both upwards and downwards, warming up the surface. This is the so-called *greenhouse effect*. The case of Venus, often called a runaway greenhouse effect, will be further developed in Section 4.2.

2.2. *Thermal escape to space*

In the homosphere and the heterosphere, collisions between molecules are frequent. Above a critical altitude, called the exobase, those collisions become less frequent, so rare that some molecules never undergo a single collision. This is the exosphere. The *exobase* is defined as the altitude above which the free mean path is comparable to or larger than the atmospheric scale height.[3] Once in this region, a molecule can escape the gravitational influence of the planet to reach space. These molecules follow ballistic trajectories determined by

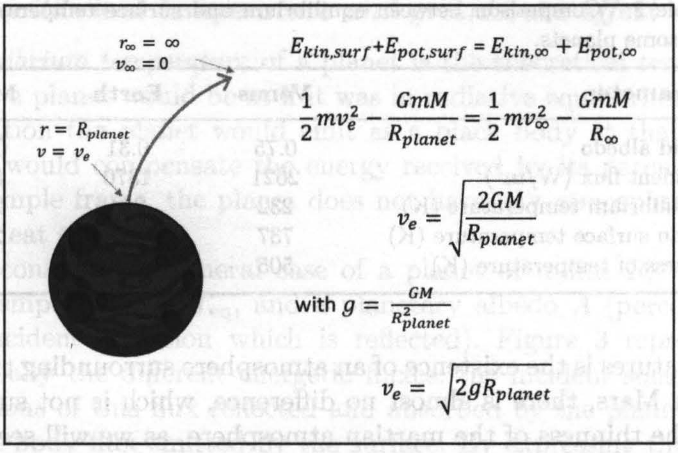

Fig. 4. Determination of the escape velocity. Here the altitude of the exobase is considered small compared to the radius of the planet.

their velocities. Most of these are captured by the gravitational field and return back to the denser regions of the atmosphere. However, some have velocities large enough to escape. The *escape velocity* (see Fig. 4) is the minimum speed needed to break free from the planet's gravity.

Whether an atmospheric molecule has sufficient speed or not depends on the temperature. As the temperature of a gas increases, its molecules move more quickly. In other words, their average speed increases. Different gases have different molecular masses and their average velocities are different at a given temperature. The velocities of individual molecules follow a Maxwellian distribution. The most probable velocity of a molecule of mass m is given by $\overline{v}_{\text{therm}} = \sqrt{2kT/m}$, where k is the Boltzmann constant. Typical values of the escape velocity and thermal velocities of H and O atoms calculated at the exobase are reported in Table 3 for different planets. The exobases of the Moon and Mercury correspond to their surface, and most of the elements ejected from their surface would be immediately lost to space. We speak of unstable or transient atmospheres. On Earth, Mars and Venus, H and He atoms can easily escape, and their atmospheres are essentially composed of molecules made of C, N and O.

Table 3. Parameters of Terrestrial Planets and typical values of the escape velocity and thermal velocities of H and O atoms for different bodies.

Parameter (unit)	Mercury	Venus	Earth	Mars	Moon
g (m/s^2)	3.7	8.8	9.8	3.7	1.62
R (km)	2439.7	6051.8	6372	3389.9	1737
v_e (km/s)	4.2	10.3	11.2	5.0	2.4
Exobase temperature (K)	725	300	1000	350	260
H $\overline{v}_{\mathrm{therm}}$ (km/s)	3.47	2.23	4.08	2.41	2.08
O $\overline{v}_{\mathrm{therm}}$ (km/s)	0.87	0.56	1.02	0.60	0.52

3. Observation Techniques

The exploration of planetary atmospheres really started with the first astronomical observations made possible by the invention of telescopes. Still today many observations are made from Earth using powerful telescopes coupled with spectrometers operating in various spectral regions. Instruments embarked on satellites also provide a rich source of information on the planetary systems they probe. In the following, we will focus on observations carried out by spectroscopic instruments on platforms orbiting the planets of interest. Figure 5 illustrates the different observation modes used to sound the atmospheres. Nadir observations correspond to an instrument looking down to the surface of the planet, during day or night. This type of measurements is sensitive to different radiative contributions, such as the radiation emitted by the planet's surface (being at a certain temperature and emitting like a blackbody), the atmosphere itself (again each layer of the atmosphere is at a certain temperature and also emits radiation), and the solar radiation being reflected by the planet. Limb observations directly sound the limb of the atmosphere investigating the emission from the atmosphere itself. Limb measurements can also be carried out when the Sun illuminates the atmosphere; in that case the signal seen by the instrument comes from solar light scattered by the atmosphere. During solar or stellar occultations the instrument points toward the Sun or a star while the Sun or the star sets down or rises.

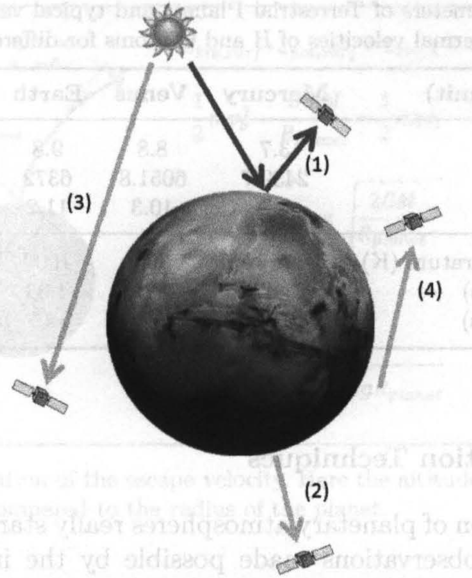

Fig. 5. Geometry of space observations: Nadir on the day side of the planet (1) or on the night side (2), solar (or stellar) occultation (3), and limb observation (4).

The theory of sounding atmospheres is based on the description of how radiation interacts with matter, in our case the atmosphere. Radiation is absorbed by the many molecules or atoms of the atmosphere. Indeed, any chemical element absorbs (or emits) radiation at specific wavelengths which are characteristic of this element. These spectral signatures cover the whole electromagnetic spectrum and correspond to transitions between two energy states of the element. In the far infrared or millimeter regions, we find essentially transitions between rotational levels, while in the mid infrared, the spectrum is due to transitions between ro-vibrational levels. In the visible and ultraviolet regions, transitions occur between electronic levels. Radiation can also be emitted by the atmosphere itself since it is lying at a certain temperature and thus radiating like any blackbody. A third mechanism of interaction is the scattering of light. Let's consider a layer in the atmosphere with infinitesimal thickness ds as schematically represented in Fig. 6, where L_λ is the incoming radiance at wavelength λ.

The attenuation of the incoming radiation after having passed through the layer, $dL_\lambda|_{\text{abs}}$, is given in Eq. (8), where β_{ext} is the

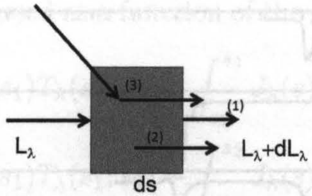

Fig. 6. Radiative transfer in the atmosphere: The incoming radiation can be absorbed (1), emitted (2). Radiation coming from different directions can also be scattered (3) in the direction of propagation considered.

extinction coefficient which depends on the physical medium and the wavelength.

$$dL_\lambda|_{\text{abs}} = -\beta_{\text{ext},\lambda} L_\lambda ds. \tag{8}$$

This coefficient encompasses the attenuation by absorption (β_a, i.e., the conversion of the energy of the radiation to heat), but also by scattering (β_s, i.e., the redirection of the incoming light out of the incoming direction of propagation):

$$\beta_{\text{ext},\lambda} = \beta_a + \beta_s. \tag{9}$$

Absorption coefficients can be calculated using first principles spectroscopy theory. They will depend on the symmetry of the molecule, the temperature, the presence of other gases. Figure 7 shows the resulting absorption spectra for a series of species usually found in atmospheres. Each molecule is characterized by specific absorption features at different wavelengths which make their detection possible with spectroscopic space instruments. The position of absorption features will give information on which molecule absorbs, and their amplitude on their abundance.

The layer of atmosphere will also emit radiation ($J_{\lambda,\text{thermal}}$) and radiation coming from different directions will be scattered in the propagation direction ($J_{\lambda,\text{scat}}$), so that the gain in radiation can be expressed by

$$dL_\lambda|_{\text{emi}} = \beta_{\text{ext},\lambda}(J_{\lambda,\text{thermal}} + J_{\lambda,\text{scat}})ds = \beta_{\text{ext},\lambda} J_\lambda ds. \tag{10}$$

Finally,

$$\frac{dL_\lambda}{\beta_{\text{ext},\lambda} ds} = -L_\lambda + J_\lambda. \tag{11}$$

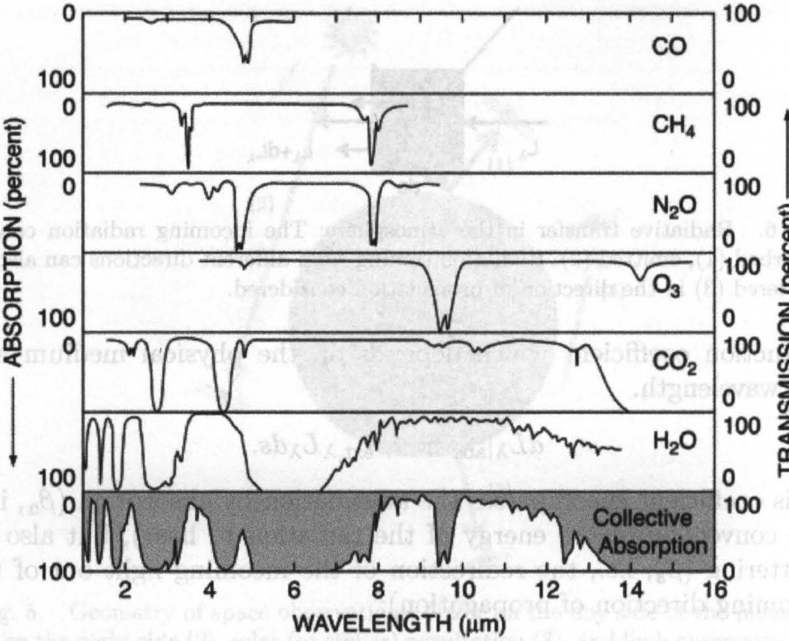

Fig. 7. Absorption spectra of several atmospheric spectra. Adapted from Ref. 2.

Introducing the *optical depth* (or the *optical thickness*) $\tau_\lambda(s_1; s_2) = \int_{s_1}^{s_2} \beta_{\text{ext},\lambda} ds$, this expression becomes:

$$\frac{dL_\lambda}{d\tau_\lambda} = -L_\lambda + J_\lambda. \tag{12}$$

This is the Schwarzchild equation, which is the fundamental description of radiative transfer summarizing the gain and loss of radiation while going through the medium. It can be written as a differential equation (Eq. (12)) or under its integral form:

$$L_\lambda(\tau_2) = L_\lambda(\tau_1) e^{-(\tau_1 - \tau_2)} - \int_{\tau_1}^{\tau_2} e^{-(\tau - \tau_2)} J_\lambda d\tau. \tag{13}$$

Let's further introduce the *transmittance* of the atmosphere along the path as $T_\lambda(\tau_1; \tau_2) = e^{-(\tau_1 - \tau_2)}$. The previous equation becomes:

$$L_\lambda(\tau_2) = L_\lambda(\tau_1) T_\lambda(\tau_1; \tau_2) + \int_{T(\tau_1, \tau_2)}^{1} J_\lambda dT, \tag{14}$$

which can also be expressed as a function of the geometric distance s:

$$L_\lambda(s_2) = L_\lambda(s_1)T_\lambda(s_1;s_2) + \int_{s_1}^{s_2} J_\lambda(s)\frac{dT_\lambda(s)}{ds}ds, \tag{15}$$

$$= L_\lambda(s_1)T_\lambda(s_1;s_2) + \int_{s_1}^{s_2} J_\lambda(s)W(s)ds, \tag{16}$$

where $W(s)$ are the weighting functions. These functions are the derivatives of the transmittance profiles $W(s) = dT_\lambda(s)/ds$. They indicate the relative contribution from a given layer of the atmosphere to the radiance received by the instrument at a given wavelength. Combining the definition of the weighting function with the definition of the transmittance and the optical depth, we see that the weighting function can also be described as the product of the transmittance by the extinction coefficient, $W(s) = \beta_{\text{ext}}T_\lambda(s)$, as illustrated in Fig. 8.

The typical form of a weighting function is a bell shape peaking at the altitude from which the measured radiation at that wavelength is originating. The radiation emitted by a layer of the atmosphere depends on the temperature of that layer, the abundances of the gases and the transmittance of the air between that layer and the measuring instrument. The radiation emitted by the lowest layers is

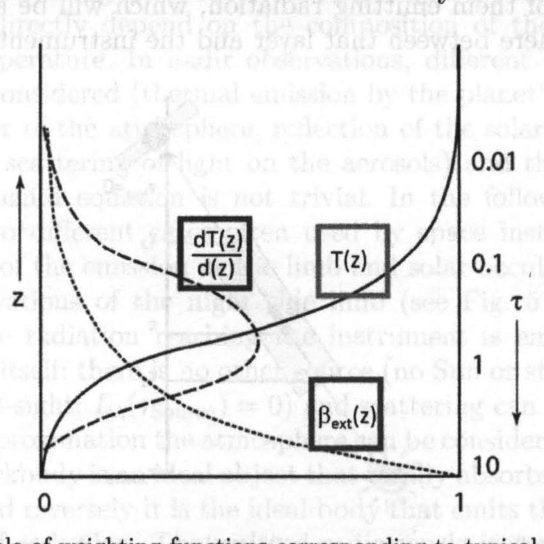

Fig. 8. Example of weighting functions corresponding to typical extinction and transmittance profiles.

high because density is high, but most of the radiation is absorbed by the atmosphere above; for the highest layers, the transmittance to space is high but the emission is low because density decreases exponentially with altitude.

Let's consider the intensity of radiation reaching an instrument (*Inst*) looking backward along the direction of propagation, which is usually the case for an instrument in space (see Figs. 5 and 9). Equation 13 can be rewritten by considering that at the level of the instrument $\tau|_{\text{Inst}} = \tau_2 = 0$ in this geometry:

$$L_\lambda|_{\text{Inst}} = L_\lambda(0) = L_\lambda(\tau_{\text{Source}})e^{-\tau_{\text{Source}}} + \int_0^{\tau_{\text{Source}}} e^{-\tau} J_\lambda d\tau. \quad (17)$$

On the left-hand side of this equation, $L_\lambda|_{\text{Inst}}$ is the intensity observed by the instrument. On the right-hand side, the first term represents the intensity of a radiation source (located at the end of the line-of-sight, LOS, of the instrument) attenuated by the atmosphere between the source and the sensor. For example, for nadir looking instrument $L_\lambda(\tau_{\text{Source}})$ would be the emission from the planet's surface and $e^{-\tau_{\text{Source}}}$ would be the total atmospheric transmittance from the surface to the instrument. The second term is the sum over all the contributions from individual infinitesimal layers along the line-of-sight, each of them emitting radiation, which will be attenuated by the atmosphere between that layer and the instrument.

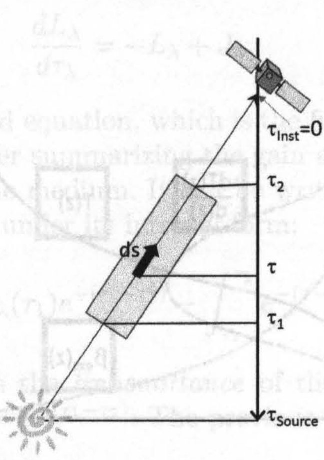

Fig. 9. Definition of the parameters for an instrument looking backward along the propagation direction.

Let's now consider scattering. When radiation is extinguished by scattering, the energy is not converted into another form, but instead redirected into another direction. The loss of radiation along the line-of-sight will be associated with a gain of radiation originating from different directions. When scattering needs to be considered, the complexity of the radiative transfer equations greatly increases, because all directions need to be taken into account into the calculations. Scattering is a complex process, as it depends on the size, form and composition of the scattering particle and on the wavelength of the incoming light. Since energy is scattered in all directions, the Schwarzschild equation becomes also more complex. As an illustration the source term J_λ of Eq. (10) becomes:

$$dL_\lambda|_{\text{emi}} = (1-\tilde{\omega})J_{\lambda,\text{thermal}} + \frac{\tilde{\omega}}{4\pi}\int_{4\pi} p(\Omega',\Omega)L_\lambda(\Omega')d\omega', \tag{18}$$

where $\tilde{\omega} = \beta_s/\beta_{ext}$ is the *single scattering albedo* of the medium. $p(\Omega',\Omega)$ is the scattering phase function, which represents the probability that a photon arriving from direction Ω' will be scattered within an infinitesimal element $d\omega$ of solid angle centered around the Ω' direction.

Radiative transfer modeling means solving the Schwarzschild equation considering the different sources and losses of radiation which will directly depend on the composition of the atmosphere and its temperature. In nadir observations, different contributions need to be considered (thermal emission by the planet's surface and by each layer of the atmosphere, reflection of the solar radiation on the surface, scattering of light on the aerosols) and the solution of the Schwarzschild equation is not trivial. In the following, we will illustrate two different cases often used by space instruments: the observation of the emission at the limb and solar occultation.

In observations of the night side limb (see Fig. 5 case [4] and Fig. 10), the radiation reaching the instrument is emitted by the atmosphere itself; there is no other source (no Sun or star at the end of the line-of-sight; $L_\lambda(\tau_{\text{Source}}) = 0$) and scattering can be neglected. As a first approximation the atmosphere can be considered as a blackbody. A blackbody is an ideal object that totally absorbs all incoming radiation and reversely it is the ideal body that emits the theoretical maximum of radiation. The emitted radiation is isotopic, and non-polarized, and only depends on the temperature of the blackbody.

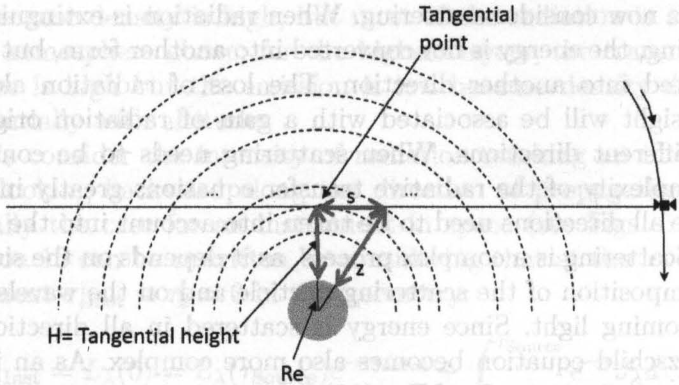

Fig. 10. Definition of the parameters for an instrument measuring the emission at the limb.

A blackbody emits more radiation than any object of the same temperature. The intensity of radiation emitted by a blackbody is given by the Planck function (Eq. (19))

$$B_\lambda(T) = \frac{2hc^2}{\lambda^5(e^{hc/k_B\lambda T} - 1)}, \tag{19}$$

where c is the speed of light, h is the Planck constant and k_B is the Boltzmann constant.

The general Schwarzchild equation reduces to

$$L_\lambda(H) = \int_\infty^0 B_\lambda(s)\frac{dT_\lambda(s;s_\infty)}{ds}ds, \tag{20}$$

where H is the tangential height, i.e., the lowest altitude sounded (see Fig. 10). Considering that the tangential height and the path along the line-of-sight (s, see Fig. 11) are linked to the altitude through:

$$(R_e + H)^2 + s^2 = (R_e + z)^2, \tag{21}$$

with R_e the radius of the planet. The expression Eq. (20) can be written as

$$L_\lambda(H) = \int_H^\infty B_\lambda(z')\frac{dT_\lambda(s;s_\infty)}{ds}\frac{ds}{dz'}dz' = \int_H^\infty B_\lambda(z')W_\lambda(H, z'_\infty)dz', \tag{22}$$

where the weighting functions are defined as

$$W_\lambda(H, z_\infty) = \begin{cases} 0 & z < H, \\ \frac{dT_\lambda(z;z_\infty)}{ds}\frac{ds}{dz} = \frac{dT_\lambda(z;z_\infty)}{ds}\sqrt{R_e/2(z-H)} & z > H. \end{cases} \tag{23}$$

The sensitivity of the instrument is increased by a factor $\sqrt{R_e/2(z-H)}$ which is due to the long tangent optical path. The weighting functions are represented for a typical observation in Fig. 11, for different tangential heights sounded sequentially as the instrument probes from the top of the atmosphere down to lower altitudes. Limb observations usually provide high vertical resolution and are sensitive to low abundances.

The second example is a particular case of limb observations, called solar occultations, which correspond to the instrument looking at the Sun. As atmospheric emission can be neglected, the equation simplifies in the known Beer Lambert law:

$$L_\lambda(H) = L_\lambda(\text{Sun})T_\lambda(\text{Sun} - \text{Inst}). \tag{24}$$

Usually solar occultation provide a very sensitive method to sound the atmosphere (long paths through the atmosphere) as well as a

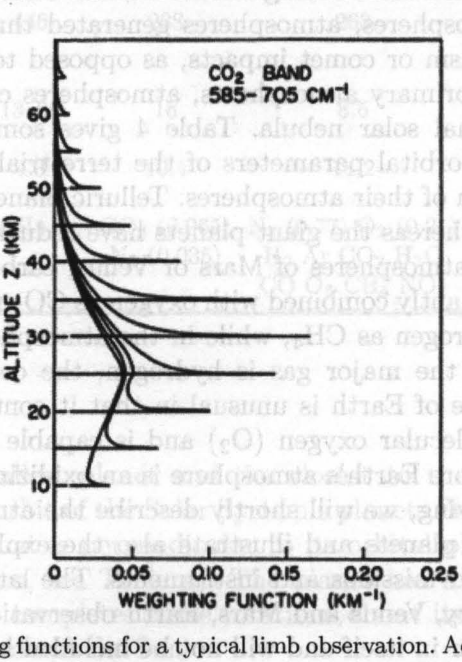

Fig. 11. Weighting functions for a typical limb observation. Adapted from Ref. 4.

high vertical resolution. With limb observations, they allow for a detailed investigation of the vertical profiles of different trace gases in the atmosphere.

In Section 4, we will describe what we know today of the atmospheres of the telluric planets of our Solar System, most of which has been obtained by spectroscopic observations carried out either from the Earth using sensitive telescopes or from space-borne instruments.

4. Terrestrial Planets

Our Solar System has four terrestrial planets: Mercury, Venus, Earth and Mars. During the formation of the Solar System, there were probably many more planetesimals, but they have all merged with or been destroyed by the four remaining worlds in the solar nebula. Only one terrestrial planet, Earth, is known to have an active hydrosphere. Telluric planets are composed mostly of some combination of hydrogen, helium and water existing in various physical states. They all have roughly the same structure: a central metallic core, mostly iron, with a surrounding silicate mantle. Those planets possess secondary atmospheres, atmospheres generated through degassing, internal volcanism or comet impacts, as opposed to the gas giants, which possess primary atmospheres, atmospheres captured directly from the original solar nebula. Table 4 gives some general information on the orbital parameters of the terrestrial planets and on the composition of their atmospheres. Telluric planets have oxidized atmospheres, whereas the giant planets have reducing atmospheres. Indeed, in the atmospheres of Mars or Venus, carbon, for example, exists predominantly combined with oxygen as CO_2 rather than combined with hydrogen as CH_4, while in the atmospheres of the giant planets, where the major gas is hydrogen, the contrary is found. The atmosphere of Earth is unusual in that it contains substantial amounts of molecular oxygen (O_2) and is capable of oxidizing the surface. Therefore Earth's atmosphere is an oxidizing one.

In the following, we will shortly describe the atmospheres of the four terrestrial planets and illustrate also the exploration of these through selected missions and instruments. The latter will be done only for Mercury, Venus and Mars, Earth observation would need a separate chapter in itself and will not be included in this section.

Table 4. Orbital parameters of terrestrial planets, as well as the composition of their atmospheres.

Parameter	Mercury	Venus	Earth	Mars
Mean distance to Sun (AU)	0.387	0.723	1.0	1.524
Mean radius (km)	2440	6051.8	6371.0	3389.9
Density (g/cm^3)	5.427	5.204	5.515	3.933
Eccentricity	0.2	0.007	0.02	0.09
Obliquity to orbit (deg)	0.034	177.4	23.4	25.2
Orbital period (days)	88.0	224.7	365.2	687.0
Rotation period (days)	58.7	−243.0	0.98	1.025
Length of day (days)	176	117	1.0	1.029
Surface temperature (K)	100–725	733	288	215
Surface pressure (bar)	0	92	1.013	0.0056
Equilibrium temperature (K)	446	238	263	222
Scale height (km)	13–95	16	8.5	18
Escape velocity (km/s)	4.3	10.4	11.2	5.0
Composition	H_2 He O_2	CO_2 (0.965) N_2 (0.035)	N_2 (0.77) O_2 (0.21) H_2 Ar CO_2 H_2O CO O_3 CH_4 NO_2	CO_2 (0.95) N2 (0.027) Ar O_2 CO H_2O

4.1. *Mercury*

Mercury, the smallest planet and the closest to the Sun, has the largest eccentric orbit of the Solar System's planets. Mercury rotates around an axis that is perpendicular to its orbital plane (its obliquity is almost zero, see Table 4). The planet has a 3:2 spin–orbit resonance, i.e., it completes three rotations about its axis for every two revolutions around the Sun.

Mercury has a significant, apparently global, magnetic field about 1.1% of the strength of Earth's. It is still strong enough to deflect the solar wind, inducing a magnetosphere. Mercury's magnetic field is distorted by the solar wind, which compresses it on the dayside and stretches it out to form a long tail on the nightside. As on Earth, it is likely that this magnetic field is generated by a dynamo effect. This dynamo effect would result from the circulation of the planet's iron-rich liquid core. Particularly strong tidal effects caused by the planet's high orbital eccentricity would serve to keep the core in the liquid state necessary for this dynamo effect.

Mercury has a tenuous atmosphere with a surface pressure of the order of or less than 10^{-12} bar. The first observations of Mercury from space were done by *Mariner 10*, which detected oxygen, helium and hydrogen atoms. Later, using ground-based instruments, other atoms, like sodium (Na), potassium (K) or calcium (Ca) were discovered. The most abundant atoms are present with number densities of the order of a few thousand atoms cm^{-3}. Sodium, potassium, oxygen and calcium atoms are thought to be surface constituents that have been extracted and blasted into the atmosphere through sputtering from the solar wind or meteoroids.[a] Indeed, Mercury's magnetic field interacts with the magnetic field of the solar wind to sometimes create intense magnetic tornadoes that funnel the fast solar wind ions down to the surface of the planet where they interact with atoms within the crust and eject them into space. Hydrogen and helium atoms are probably captured from the solar wind, of which they are major constituents. This tenuous atmosphere is not stable. It is continuously lost to space and replenished by the solar wind or impacts. Indeed escape to space is easy because of the high temperatures reigning on the day side. Along with its weak gravity field, it means that for almost any atom or molecule, their thermal velocity will be larger than the escape velocity (about 4 km/s, see Table 3). In fact, the exosphere, i.e., the upper layer of the atmosphere which gradually fades into space, starts at its surface. The huge temperature

[a]When a fast atom or ion hits an atmospheric atom, the latter gains energy and may be able to escape the planet's gravitational attraction. In the case of tenuous atmospheres, the fast ion or atom hits the surface directly, ejecting one or several atoms from the crust into space.

difference between day (725 K) and night (80 K) comes from the fact that the planet rotates very slowly (a day on Mercury lasts 176 Earth days) and the atmosphere is so thin it does not retain heat.

The surface is rocky with a multitude of craters, indicating that the planet has been geologically inactive for billions of years. There are no signs of volcanic activity or plate tectonics. The high number of craters on the surface is explained by its tenuous atmosphere.

Mariner 10 was the first mission to use the gravity of one planet (in this case Venus) to reach another one. The spacecraft was launched in November 1973 toward Venus, and after a flyby of that planet, its trajectory was modified toward Mercury which was reached in March 1974. *Mariner 10* was then set to orbit the Sun, flying above Mercury again in September 1974. A third flyby occurred in March 1975 at an altitude of 305 km. In total *Mariner 10* took pictures of about half Mercury's surface. *MESSENGER* (Mercury Surface, Space Environment, Geochemistry and Ranging[5]), the next mission to Mercury, was launched in 2004, so about 30 years after *Mariner 10*. The spacecraft entered orbit around Mercury in March 2011, and started acquiring data soon after. The mission discovered high concentrations of magnesium and calcium on the dark side of the planet,[6,7] identified a significant northward offset of the planet's magnetic field, found large amounts of water in the exosphere, evidence of water ice at the poles, frozen in locations that never see the sunlight,[8,9] and revealed evidence of past volcanic activity on the surface. Beginning of the summer 2014, the orbit was lowered down to 25 km altitude to be able to carry out new research. However it was soon found out that the spacecraft propellant was running out and that it would soon impact the planet. As expected, on April 30, 2015, *MESSENGER* slammed into Mercury's surface creating a new crater on the planet. *BepiColombo*,[10] the first European mission to Mercury, has been launched in October 2018 for a 7-year journey to the planet. It comprises two spacecraft: the Mercury Planetary Orbiter (MPO) and the Mercury Magnetospheric Orbiter (Mio). BepiColombo is a joint mission between the European Space Agency (ESA) and the Japan Aerospace Exploration Agency (JAXA). MPO will investigate the surface and internal composition of the planet; Mio will study its magnetosphere and magnetic field. As of today, *BepiColombo* has successfully performed its first flyby of the planet (see Fig. 12). *BepiColombo*'s main science mission will begin in early 2026.

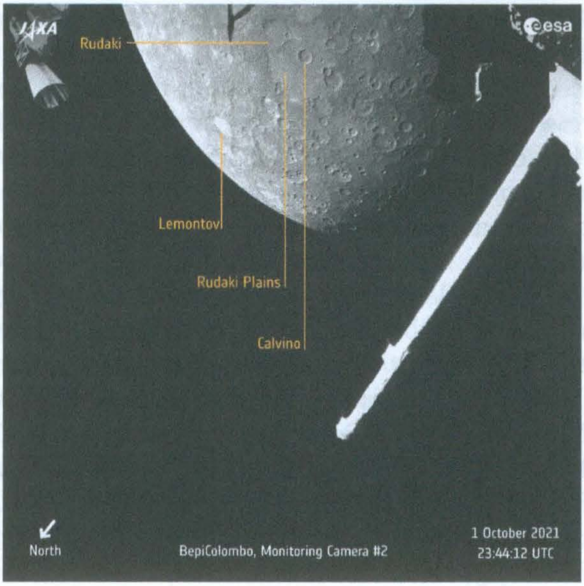

Fig. 12. The joint European-Japanese *BepiColombo* mission captured this view of Mercury on October 1, 2021 as the spacecraft flew past the planet for a gravity assist manoeuvre. The region shown is part of Mercury's northern hemisphere. Credit: ESA/BepiColombo/MTM, CC BY-SA 3.0 IGO.

4.2. *Venus*

Venus has an atmosphere very different from that of Earth. In comparison to Earth it is much denser, heavier and extends to a much higher altitude. The temperature and pressure at the surface are 740 K and 92 atm, respectively. Despite the harsh conditions on the surface, at about a 50-km to 65 km level above the surface of the planet, the atmospheric pressure and temperature are nearly the same as those on Earth's surface.

The atmosphere of Venus is mostly made up of CO_2 (97%) and nitrogen (N_2, less than 3.5%), the rest being trace gases such as carbon monoxide (CO), water vapor (H_2O), halides (HF, HCl), sulfur-bearing species (SO_2, SO, OCS, H_2S) and noble gases. Sulfur compounds are extremely important to understand the formation of the Venusian clouds which are believed to be composed of sulfuric acid (H_2SO_4) droplets. These clouds completely enshroud the planet in a series of layers, extending from 50 to 70 km altitude, and are composed of particles of different sizes and different H_2SO_2/H_4O compositions.[11] These act also as a very effective separator between the atmospheres below and above the clouds, which show very distinctive characteristics. Water vapor is present in extremely low quantities (about 30 parts per million or ppm, 1 ppm = 0.001%), making Venus the driest planet of the solar system.[12,13]

As on Earth, the atmosphere can be divided into distinct regions based on the evolution of the temperature with altitude, as shown in Fig. 13. The troposphere on Venus extends from the surface up to about 70 km altitude which also corresponds to the top of the main cloud deck. No temperature variations (diurnal or seasonal) have been observed below the clouds. Above the clouds, the mesosphere extends up to about 100 km. At higher altitudes, in the thermosphere, very different temperatures for the night and day sides have been observed. The nightside is so cold that some are referring to it as the cryosphere. Due to the absence of a magnetosphere, the solar wind interacts directly with the ionosphere, confining the latter below 200 km altitude.

The dynamics of the Venusian atmosphere is quite complex and characterized by two regimes of circulation: *retrograde zonal superrotation* in the troposphere and mesosphere and *solar-antisolar circulation* in the thermosphere. The zonal circulation is characterized by wind speeds decreasing from 100 m/s at the cloud top to

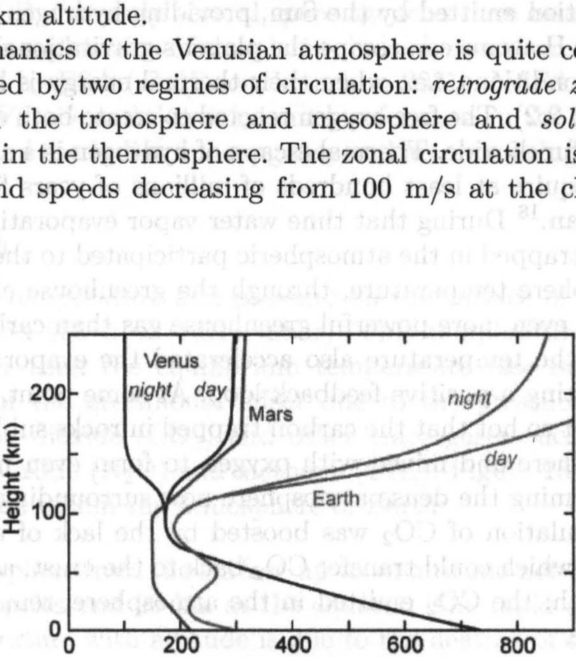

Fig. 13. Variation of the temperatures within the atmospheres of Venus, Earth and Mars.

almost 0 at the surface. The mechanisms that maintain the superrotation are still not fully understood.[14]

The equilibrium temperature of Venus is only about 240 K, whereas the temperature at the surface reaches much higher values (see Table 4). This difference is attributed to the huge greenhouse effect experienced by the planet. Venus and Earth formed from the same interstellar gas and dust, and therefore their initial composition should be very similar.[15] Liquid water has always been present at the surface of Earth, while Venus evolved toward a dry and hot planet. Some proposed that Venus could have been formed in a drier region of the Solar nebulae which would explain the current depletion in water.[16] However, observations indicate that, in its past, Venus had more water than in the present epoch. This is evidenced by measurements of present-day D/H, which indicate an overabundance of deuterium (heavy hydrogen, D) relative to hydrogen. Venus might even have supported enough water to form an ocean. Over time, this water was vaporized[17,18] then lost through photodissociation by ultraviolet (UV) radiation emitted by the Sun, providing hydrogen and oxygen atoms. The H atoms can escape the planet's gravitational field under favorable conditions, i.e., when their thermal energy is high enough (see Section 2.2). The free oxygen reacted to create both carbon dioxide and sulfur dioxide. Thermal escape of hydrogen is a slow process; it would require at least hundreds of millions of years for Venus to lose its ocean.[18] During that time water vapor evaporating from the ocean and trapped in the atmospheric participated to the increase of the atmosphere temperature, through the greenhouse effect. Water vapor is an even more powerful greenhouse gas than carbon dioxide. Increasing the temperature also accelerated the evaporation of the ocean, creating a positive feedback loop. At some point, the surface of Venus got so hot that the carbon trapped in rocks sublimated into the atmosphere and mixed with oxygen to form even more carbon dioxide, forming the dense atmosphere now surrounding the planet. The accumulation of CO_2 was boosted by the lack of any efficient CO_2 cycle, which could transfer CO_2 back to the crust, as is happening on Earth: the CO_2 emitted in the atmosphere, remained in the atmosphere.

The composition and photochemical processes occurring below, within and above the clouds are quite different in nature. The chemistry in Venus's atmosphere is controlled by ion–neutral and ion–ion reactions in the ionosphere, by photochemistry within and above the cloud global layer, and by thermal equilibrium chemistry, which prevails near the surface.[19] Venus's main cloud layers span the 50–70-km altitude range, with hazes reaching up to 90 km and down to 30 km. The main cloud deck is composed of different populations of particle sizes ranging from less than a micrometer up to 35 μm.

Although Lomonosov was the first to suggest that Venus had an atmosphere, through observations of the 1761 Venus transit, the systematic observation of Venus started in the early 20th century with the detection of the main constituent of the atmosphere, CO_2, the determination of the cloud structure, the high surface temperature and the existence of the greenhouse effect gathered from Earth-based observations and the first fly-by of the planet by *Mariner 2*. This was followed by a long series of space missions, including the *Venera* and *Mariner* missions, *Pioneer Venus* or *Magellan* (see Ref. 20 for a review) and more recently *Venus Express* of the European Space Agency[21,22] and *Akatsuki*,[23] which is the last mission still orbiting the planet today. But already several space agencies have been supporting the future exploration of the planet, indeed several missions have been selected by ESA (*EnVision*, launch in 2031), NASA (with two missions *DaVinci+* and *Veritas*, launches foreseen in the 2028–2030 timeframe) and the Indian space agency and are now being prepared.

4.3. *Earth*

The average temperature and pressure on the surface of the Earth are 288 K and 985 hPa (1,013 mbar). This temperature is about 35 K warmer than the equilibrium temperature (see Section 2.1) as a result of the greenhouse effect due to the presence of water vapor, carbon dioxide (CO_2) and other trace gases such as ozone (O_3), nitrous oxide (N_2O) and methane (CH_4). Figure 13 shows the different layers within the atmosphere of Earth:

- The troposphere from the surface up to an altitude between 10 km above the poles and 20 km at the equator. The rapid decrease of the temperature with altitude is due to the heating of the surface and causes convection.
- The stratosphere, characterized by an increase of temperature with altitude because of the presence of ozone, whose abundance is the

highest in the stratosphere, forming what is called the *ozone layer*. Ozone is the only component that significantly absorbs the UV solar radiation penetrating the atmosphere. Despite its low concentration, ozone effectively absorbs the solar radiation between 230 and 300 nm, protecting the Earth against harmful ultraviolet radiation which therefore cannot reach the lower layers of the atmosphere, nor the surface. The energy absorbed by ozone is then transferred to kinetic energy, heating the stratosphere. This region is unique to Earth, and is not found in the other atmospheres of the telluric planets.

- The mesosphere, in which the temperature decreases with altitude, caused by a decreasing production of ozone and an increase in the cooling rate due to CO_2.
- The thermosphere, above 80–90 km, where temperature increases with altitude, following the absorption of the UV solar radiation and a rarefied atmosphere which prevents any efficient cooling through IR emission.

The atmosphere of Earth primarily consists of nitrogen (N_2, 78%) and molecular oxygen (O_2, 21%), and a series of trace gases, including water (H_2O), argon (Ar) and CO_2 (see Table 4). The main characteristics of the Earth and difference compared to the other planets of our Solar System is the presence of liquid water on its surface. It is likely that the presence of liquid oceans played an essential role in the development of life on Earth. The appearance of life also had irreversible impacts on the atmosphere, with the production of oxygen, whose photodissociation led to the formation of the ozone layer.

4.4. *Mars*

Mars lost its magnetosphere 4 billion years ago, so the solar wind interacts directly with the Martian ionosphere, keeping the atmosphere thinner than it would otherwise be by stripping away atoms from the outer layer. Both *Mars Global Surveyor* and *Mars Express* have detected these ionized atmospheric particles trailing off into space behind Mars.[24] Atmospheric pressure on the surface varies from around 30 Pa on Olympus Mons to over 1,155 Pa in the depths of Hellas Planitia, with a mean surface level pressure of 600 Pa. This is less than 1% of the surface pressure on Earth. The equivalent pressure of Mars surface is found at a height of 35 km above the Earth's surface. The scale height of the atmosphere is about 11 km; higher than Earth's 6 km due to the lower gravity. The average surface temperature is about 215 K, although large latitudinal, seasonal and diurnal variations are observed. For example, the temperature at the equator drops to 200 K at night and reaches 300 K during the day.

The atmosphere on Mars consists of 95% carbon dioxide, 3% nitrogen, 1.6% argon and contains traces of oxygen and water. Liquid water cannot exist on the surface under the current temperature and pressure conditions, although the surface of the planet is covered by many evidences of liquid water, suggesting a warmer and wetter past.

Mars surface is mostly composed of iron-rich basaltic and andesitic rock. Surface dust, also called regolith, is ubiquitous on the planet and can be lifted up in the atmosphere. The presence of hematite (Fe_2O_3) and goethite (α-FeO(OH)) explains the reddish color of the planet. Because there is no plate tectonics taking place on Mars, the surface preserves traces of its past. Scientists could estimate the age of different areas on the planet, based on the number and size of craters, as far as 4.5 billion years ago, while on Earth the surface is younger than 2 billion years. Although Mars does not have any global magnetic field, observations revealed traces of remnant magnetic field[25] resulting from the magnetization of some terrains in the past of the planet.

Mars has an elliptical orbit with perihelion and aphelion of, respectively, 1.38 and 1.66 AU. The complete revolution around the Sun takes 687 terrestrial days. The rotation of the planet on itself, i.e., a martian day (also called "sol") lasts a little longer than Earth's, about 24 h 40 min. Obliquity is about 25°, more or less the same as Earth's, implying that Mars has also a seasonal cycle. But due to its elliptical orbit, seasons on Mars differ more between northern and southern hemisphere than on Earth. On Mars, the *Solar Longitude*, L_S, is used to define the seasons and more generally the time (see Fig. 14). L_S ranges from 0° to 360° corresponding to a complete revolution around the Sun. L_S represents thus the angle between Mars, the Sun and a reference position of Mars at $L_S = 0°$ corresponding to the spring equinox in the northern hemisphere. $L_S = 90°$

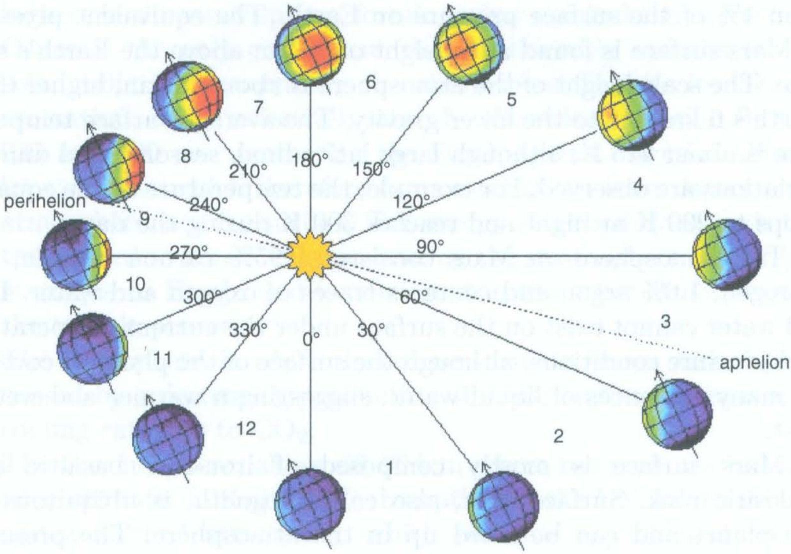

Fig. 14. Solar longitude on Mars. Adapted from Ref. 26.

corresponds then to the northern summer solstice, $L_S = 180°$ to the northern autumn equinox and $L_S = 270°$ to the northern winter solstice. The first Martian year (MY1) was arbitrarily chosen to start at the northern spring equinox ($L_S = 90°$) on April 11, 1955.[27]

Because the orbit of Mars is highly elliptical, the orbital revolution velocity is the highest while the planet is closest to the Sun and Mars receives up to 45% more energy than at aphelion. As a consequence, the duration and strength of the seasons in both hemispheres are quite different. The southern hemisphere undergoes more extreme seasons with a warm short summers and a cold long winters, while in the northern hemisphere, the seasons are more temperate with longer summers but less warm and shorter winters with less cold temperatures.

Like Earth, Mars possesses polar caps whose extensions toward lower latitudes vary throughout the year. On Mars, the polar caps are characterized by a residual (present all year round) and a seasonal cap (appearing during the cold seasons). At both poles, a mix of sediments, dust and ice has accumulated over time to form a deposit, extending on about 1,000 km large and whose thickness reaches more than thousands of meters. These polar deposits are partly covered by the residual polar caps. The residual caps at the north and south poles differ in composition and size[28]: in the north, the deposits are covered by a layer of water ice while in the south, the cap is smaller and is made of CO_2 ice (see Fig. 15). As the temperature decreases, atmospheric CO_2 condenses creating the seasonal cap, which is released back to the atmosphere when the temperature rises again. This leads to important variations of the atmospheric pressure.

The atmosphere of Mars is constantly loaded with dust whose abundances depend on season and location.[29] The amount of dust in the atmosphere of Mars has a seasonal cycle (see Fig. 16), with the first half of the year having less dust than the second half (in the northern hemisphere). During summer in the southern hemisphere, Mars is also at perihelion in its orbit around the Sun, which causes an increase in the amount of dust lifted from the surface and suspended in the atmosphere. There is large inter-annual variability in the dust activity, with some years having smaller regional scale dust storms and some years developing into a global, planet-encircling dust storm. The nature of this variability is still unknown and is an open question in Mars research. The dust and water cycles[30] are coupled through cloud condensation processes, but dust also modifies the thermal

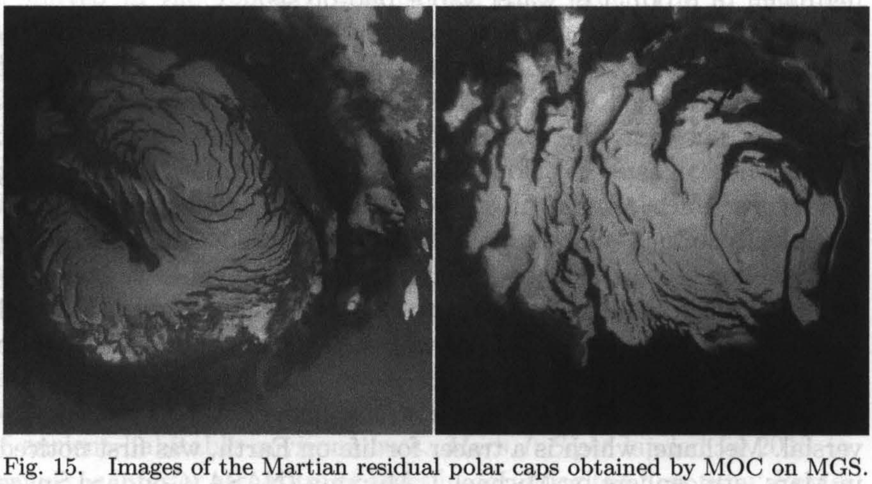

Fig. 15. Images of the Martian residual polar caps obtained by MOC on MGS. (Left) North polar cap (dimensions: roughly 1,100 km across); (Right) South polar cap (dimensions: from right to left, about 420 km across). Credit: NASA/JPL/MSSS.

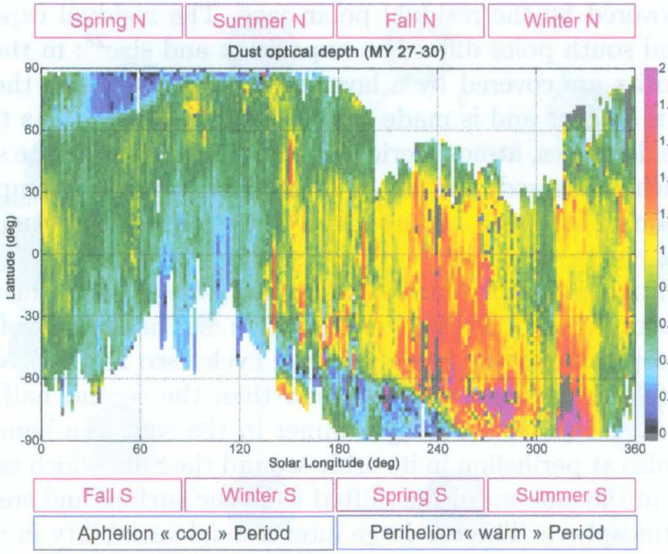

Fig. 16. Dust climatology. Adapted from Ref. 31.

structure of the atmosphere. Recent studies suggest that global dust storms effectively transport water vapor from the near-surface to the middle atmosphere and increase the escape rate of atmospheric hydrogen (a product of water vapor photolysis).[32,33]

Different kinds of clouds occur on Mars: water ice and CO_2 ice clouds.[34] Water ice clouds usually appear under situations of adiabatically cooled upward flows. The dust particles serve as condensation nuclei in the formation of the clouds. However, as dust warms the atmosphere, clouds occur only under low dust loading conditions. Such clouds can take several forms: polar hoods, topographically induced clouds, cloud belt forming at low latitudes during the aphelion season. CO_2 ice clouds can only form when the temperature is low enough to allow CO_2 to condense. Such clouds are found in the polar regions during the winter. High altitude mesospheric clouds have also been observed.[35]

The presence of methane in the atmosphere of Mars is still controversial. Methane, which is a tracer for life on Earth, was first noticed in Mars' atmosphere by Michael J. Mumma (NASA/Goddard Space Flight Center), whose team first detected Martian methane in 2003.[36] In March 2004, Mars Express confirmed the presence of methane in the Martian atmosphere with a concentration of about 10 ppb by volume.[37,38] Since methane is an unstable gas that is broken down by ultraviolet radiation, typically lasting in the atmosphere for about 340 years, its presence on Mars could indicate that there is (or has been within the last few hundred years) a source of the gas on the planet. Volcanic activity, comet impacts, and the existence of life in the form of microorganisms such as methanogens have been proposed as possible sources, although some are now thought not to have been important.[39] Methane appears to occur in patches, which suggests that it is being rapidly broken down before it has time to become uniformly distributed in the atmosphere, and so it is presumably also continually being released to the atmosphere. It has also been shown that methane could also be produced by a non-biological process involving water, carbon dioxide, and the mineral olivine, which is known to be common on Mars.[40] One of the main science objectives of the ExoMars Trace Gas Orbiter mission is to confirm or refute the presence of the trace gas within the atmosphere of Mars. Since the beginning of its science phase in April 2016, methane has been continuously monitored by both spectrometers on board the spacecraft, i.e., NOMAD and ACS. Up to now, no detection of methane could be reported.[41,42]

The history of the exploration of the Red Planet is very rich; it started in the 1960s. Following the successful flyby in 1965 by *Mariner 4*, several missions were designed and launched by the Soviet Union and the USA with various success rates. Today six space agencies have successfully made it to Mars: NASA, the former Soviet Union space program, the European Space Agency (*Mars Express*, *ExoMars Trace Gas Orbiter*), the Indian Space Research Organization (*MOM*), the China National Space Administration (*Tianwen-1*), and the United Arab Emirates Space Agency (*Hope*). Today several spacecraft orbit around the planet continuing to deliver a wealth of information on its atmosphere, helped by a series of rovers crossing over the surface of the planet delivering ground truth about the surface composition and the lower layers of the atmosphere. NASA's rover *Curiosity*, arrived at Gale Crater in 2012 to search for signs of ancient habitable environments. *Mars InSight* landed in 2018 to probe the interior structure of Mars in detail for the first time. *Perseverance*, which landed on Mars in Feb. 2021, will search for potential signs of life. *Perseverance* also carries a test helicopter, *Ingenuity*, which will assess the feasibility of flying on Mars. *Perseverance* will cache the most promising samples for a future sample-return mission, tentatively scheduled for later in the decade and involving both NASA and the European Space Agency. In the near future, the second element of the *ExoMars* mission will deliver a rover and a platform on the surface of Mars (launch expected in 2022). Japan is now building the Mars Moons Exploration (*MMX*) mission for a sample-return mission from Phobos, one of the two moons of Mars.

5. Q&A

Julia Maia: You said that indications of volcanism on Venus include SO_2 and high emissivity surface rocks. What else could cause high emissivity? What other data could test that?

Ann-Carine Vandaele: New missions are needed to study the emissivity problem. The high emissivity could be an artifact of higher temperature caused by internal heat, but could also result from compositional and even structural (e.g., particle size) differences. Multiple signs of volcanic activity are seen in the geology of the surface, but there is no direct evidence of current activity.

Luisa M Lara: Volcanoes emit more than SO_2. If Venus is active, future missions could detect these other gases?

Ann-Carine Vandaele: Water and carbon monoxide are two such gases that could be measured spectroscopically. We are planning a suitable spectrometer for next European mission to Venus.

Maya Garcia-Comas: Why is there hemispheric asymmetry (more in the North) in the amount of polar cloud on Mars?

Ann-Carine Vandaele: The orbit of the planet is quite eccentric, and Mars has a non-zero obliquity. So there is a seasonal effect that causes the asymmetry because one pole is pointing nearer the Sun at perihelion while the other is not. We do not notice this effect on Earth because, although we have an obliquity, the orbital eccentricity is very small.

Maya Garcia-Comas: What are the conditions for a regional storm to become a global storm. Why do these storms start and stop?

Ann-Carine Vandaele: It is still not really understood. The main idea is that it is somehow related to the orbit, but it is not clear. There may also be a kind of runaway once a lot of dust gets in the atmosphere and changes the thermal structure. But we don't know.

Julia Maia: Obs Cote d'Azur: Why is Mars so heavily explored yet Venus left aside?

Ann-Carine Vandaele: Venus is easier to reach but harder to study, because the atmosphere is hotter, the pressure is higher and the cloud cover is total. Also, human exploration of Venus has no future, while human exploration of Mars has been on people's minds for decades, or even centuries.

Asier: How do you get vertical resolution from a limb profile of an atmosphere?

Ann-Carine Vandaele: You have many lines of sight through the atmosphere each with different impact parameters, so you can take differences to probe the change between adjacent lines, like peeling an onion.

Olga Muñoz: Can you explain about the relation between dust in Mars' atmosphere and water escape?

Ann-Carine Vandaele: We see that there is more water at high altitudes and faster escape of hydrogen when the atmosphere is dusty. This is because the dust warms the atmosphere by absorbing sunlight and then the atmosphere expands upwards, exposing more water to the solar UV flux and dissociation. So, it's nothing much to do with reactions, but with a change in the atmospheric structure.

Mayer: How do you get to be a PI of a space instrument and how do you have time for science when you are a PI?

Ann-Carine Vandaele: Like what you do, build your luck by participating, by networking, to show that you do good work, and have some luck.

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