

Review



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Possible climates on terrestrial exoplanets

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What kind of environment may exist on terrestrial planets around other stars? In spite of the lack of direct observations, it may not be premature to speculate on exoplanetary climates, for instance, to optimize future telescopic observations or to assess the probability of habitable worlds. To begin with, climate primarily depends on (i) the atmospheric composition and the volatile inventory; (ii) the incident stellar flux; and (iii) the tidal evolution of the planetary spin, which can notably lock a planet with a permanent night side. The atmospheric composition and mass depends on complex processes, which are difficult to model: origins of volatiles, atmospheric escape, geochemistry, photochemistry, etc. We discuss physical constraints, which can help us to speculate on the possible type of atmosphere, depending on the planet size, its final distance from its star and the star type. Assuming that the atmosphere is known, the possible climates can be explored using global climate models analogous to the ones developed to simulate the Earth as well as the other telluric atmospheres in the solar system. Our experience with Mars, Titan and Venus suggests that realistic climate simulators can be developed by combining components, such as a 'dynamical core', a radiative transfer solver, a parametrization of subgrid-scale turbulence and convection, a thermal ground model and a volatile phase change code. On this basis, we can aspire to build reliable climate predictors for exoplanets. However, whatever the accuracy of the models, predicting the actual climate regime on a specific planet will remain challenging because climate systems are affected by strong positive feedbacks. They can drive planets with very similar forcing and volatile inventory to completely different states. For instance, the coupling among temperature, volatile phase changes and radiative properties results in instabilities, such as runaway glaciations and runaway greenhouse effect.

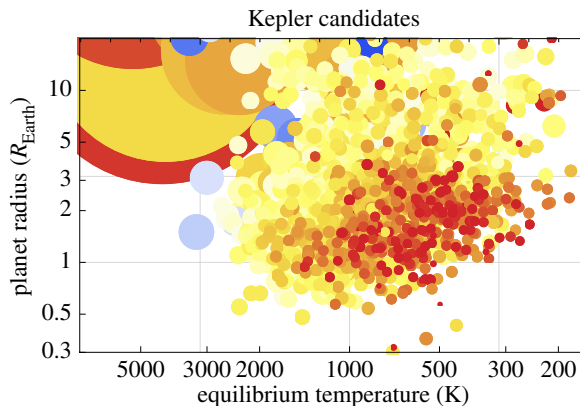


Figure 1. Kepler planet candidates in a radius-equilibrium temperature diagram. The size and the colour of each dot are, respectively, representative of the size and colour of the parent star. This diagram suggests the absence of any gap in the planet radius distribution between Earth- and Neptune-size planets. The ‘equilibrium temperature’ T_e is obtained assuming a planetary albedo set to zero: $T_e = (F/4\sigma)^{0.25}$, with F the mean stellar flux at the orbital distance and σ the Stefan–Boltzmann constant ($\sigma = 5.6710^{-8}$ SI). (Online version in colour.)

1. Introduction

To help in designing future ground-based or space telescopes aiming at characterizing the environment on terrestrial exoplanets or to address scientific questions like the probability of habitable worlds in the galaxy, one has to make assumptions on the possible climates and atmospheres that may exist on terrestrial exoplanets. For this, speculation is unavoidable because no direct observations of terrestrial atmospheres are available outside the solar system. The limited sample that we can observe here suggests that a wide diversity of planetary environment is possible. Would we imagine Venus or Titan if they were not there?

Fortunately, observational statistics on the exoplanets themselves are starting to be available. The extrapolation of super-Earth detections suggests that terrestrial planets should be abundant in our galaxy. A large fraction of the stars are likely to harbour rocky planets [1–4]. These discoveries have also profoundly changed our vision of the formation, structure and composition of low-mass planets: while it has been long thought, mostly based on the observations of our own Solar System, that there should be a gap between telluric planets with a thin, if any, secondary atmosphere and the so-called icy giants that retained a substantial amount of hydrogen and helium accreted from the protoplanetary disc, this gap does not seem to exist in exoplanetary systems.

As can be seen in figure 1, the distribution of the radius of planet candidates detected by the Kepler space telescope [5] is quite continuous from 0.7 up to 10 Earth radii, and particularly between 2 and 4 Earth radii, where the transition from Earth- to Neptune-like planets was thought to occur. Although these observations are still incomplete—especially when planets get smaller, have a lower equilibrium temperature, or orbit bigger stars—they suggest that there may not be a clear-cut distinction between low-mass terrestrial planets and more massive planets for which the gaseous envelope represents a significant fraction of the bulk mass. If such a continuum exists in the bulk composition of low-mass planets, one can also anticipate that the various atmospheric compositions seen in the Solar System are only particular outcomes of the continuum of possible atmospheres.

This raises several pending questions. What kind of atmospheres can we expect? Can we relate the global, measurable parameters of a planet (mass, radius, intensity and spectral distribution of the incoming stellar energy, etc.) to the mass and composition of its atmosphere, and ultimately predict a range of possible climates?

In this paper, written for non-specialists, we review the different processes which may control the environment on terrestrial exoplanets, including their habitability. In §2, we speculate on the possible diversity of atmospheric composition and mass, which depends on complex processes, which are not easy to model: origins of volatiles, atmospheric escape, geochemistry, long-term photochemistry. In §3, we mention the importance of the body rotation (period and obliquity) and evaluate the impact of gravitational tides on planetary spin. If the atmosphere and the rotation are known, we explain in §4 that, the corresponding possible climates can be explored based on the likely assumption that they are controlled by the same type of physical processes at work on solar system bodies. In particular, this can be achieved using models analogous to the ones developed to successfully simulate the Earth climate as well as Mars, Venus, Titan, Triton, Pluto. However, as discussed in §5, whatever the accuracy of the models, predicting the actual climate regime on a specific planet will remain challenging, because climate systems are affected by strong positive feedbacks (e.g. coupling among radiative properties, temperatures and volatiles phase changes) and instabilities which can drive planets subjected to very similar volatiles inventory and forcing to completely different states.

2. Which atmosphere on exoplanets?

(a) Origins of atmospheres

To understand the various possible types of atmospheres, one first needs to consider the various sources of volatiles available during the formation of the planet. These sources have mainly two origins: the nebular gas present in the protoplanetary disc during the first 1–10 Myr of the planet formation and the volatiles (mainly H₂O and CO₂) condensed and trapped into the planetesimals accreting on the nascent planet (and possibly into the comets or asteroids colliding with the planet after its formation in the so-called ‘late veneer scenario’). The volatiles initially incorporated in the bulk of the mantle can be released through two major channels, catastrophic outgassing and release by volcanism, with very different timescales.

As discussed in §2*b*, because atmospheric escape is closely related to stellar activity, it is strongly time-dependent at early ages. The timescale on which the various species can be added to the atmosphere is thus critical in determining what is left in the matured atmosphere. Hence, we discuss these three formation channels and their associated timescales separately.

(i) Nebular gas and protoatmospheres

When a dense, cold molecular cloud gravitationally collapses to form a protostar, conservation of angular momentum forces a fraction of the gas to remain in an extended disc where planets can form. This gas is mainly composed of hydrogen and helium. The abundances of heavier elements are expected to be close to the stellar ones, except for some elements that can be trapped in condensing molecules. While these discs may be quickly dispersed by stellar radiation and winds (on timescales of the order of 3 Myr [6]), planetary embryos more massive than approximately 0.1 M_{Earth} can retain a significant mass of nebular gas, depending of course on the local conditions in the nebula, on the core mass and on the accretion luminosity [7,8].

An extreme case occurs when the embryo becomes massive enough and the mass of the atmosphere becomes similar to the core mass. Then, the so-called core instability can be triggered, resulting in an unstable gas accretion that can proceed almost until all the available gas in this region of the disc is fed to the planet [9–11]. This is the mechanism that is thought to have formed the four giant planets in our Solar System. The critical core mass above which the core instability is triggered could be as low as a few Earth masses, and a substantial primitive atmosphere could be accreted by much smaller planets [7].

This is illustrated in figure 2, where the masses and radii of all the observed low-mass transiting planets so far are reported. If an object exhibits a radius that is bigger than the radius that would have a body composed entirely of water (water being the least dense of most abundant

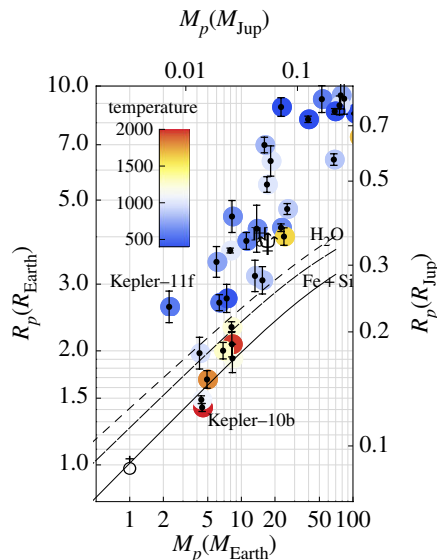


Figure 2. Mass-radius diagram of planets in the Earth to Saturn mass regime. The colour of each dot is related to the equilibrium temperature of the planet (see colour bar in K). Curves represent the mass radius relationship for an Earth-like planet with a water mass fraction of 0, 0.5 and 1 from bottom to top. Planets above the top curve must have a massive gaseous envelope to explain their large radii. One can see that in the low-mass regime, hotter planets preferentially have a higher density that is either due to the more efficient escape or to lower gas accretion efficiency in hot regions of the disc. ‘Temperature’ corresponds to the planetary equilibrium temperature, as in figure 1. (Online version in colour.)

material except for H/He) of the same mass (dashed curve), this tells us that at least a few per cent of the total mass of the planet are made of low-density species, most likely H_2 and He gas. The fact that many objects less massive than Neptune are in this regime indeed confirms the possibility to accrete a large fraction of gas down to $2\text{--}3 M_{\text{Earth}}$, the mass of Kepler-11 f.

The fact that, in a given mass range, radii can easily vary by a factor of two reminds us, if need be, that the early gas accretion depends on many parameters that are not well understood (mass and dispersal time of the disc that can change from one system to another, location of the protoplanet, etc.). The determination of the gas mass fraction of a given object, even knowing its current properties, is far from being a trivial task.

(ii) Catastrophically outgassed $\text{H}_2\text{O}/\text{CO}_2$ atmospheres

The other source of volatiles are the planetesimals that accrete to form the bulk of the planet itself. These will be the major sources of (i) carbon compounds like CO_2 or possibly CH_4 , (ii) water, especially if they formed beyond the ‘snow line’ (the distance from the star in the nebula where it is cold enough for water to condense into ice grains), and (iii) to a lesser extent N_2/NH_3 and other trace gases.

In current terrestrial planet formation models, planets are usually formed in less than 100 Myr. During this phase, the energy produced by the impacts of the planetesimals and planetary embryos is generally large enough to melt the upper mantle, creating a planet-wide magma ocean. When the accretion luminosity decreases, however, this magma ocean starts to solidify. Because solidification is more easily initiated at high pressures and molten magma is less dense than the solid phase, this solidification proceeds from the bottom upward [12]. During this phase, which can last from 10^5 to 3×10^6 years depending on the volatiles fraction, H_2O and CO_2 , which cannot be trapped in the solid phase in large quantities, are rapidly outgassed.

The mass of the resulting atmosphere then depends on the composition of the planetesimals, and thus on their initial location as well as on the metallicity of the star.

For a planet like the Earth formed at warm temperatures (where water ice is not stable), the available amount of volatiles should be limited because the water mass fraction in the planetesimal should be low. It is estimated that no more than a few Earth oceans equivalent mass ($EO \approx 1.4 \times 10^{21}$ kg \approx 270 bar on the Earth) of water and 50–70 bar of CO_2 were released that way ([13] and references therein).

If the planet is formed much closer to, or even beyond, the snow line, the water content of the planetesimals could be much larger (a few 10 wt%), and tens to thousands of Earth oceans of water could be accreted [14]. This suggests the existence of a vast population of planets with deep oceans (aqua-planets) or even whose bulk composition is dominated by water (ocean planets; [15]). In that case, the physical state of the outer water layer (supercritical, steam, liquid water, ice) depends on the temperature that is first controlled by the cooling mantle during the first tens of million years, and then by the insolation received.

(iii) Volcanically degassed secondary atmospheres

On a much longer, geological timescale, the volatiles that remained trapped in the mantle during the solidification can be released through volcanic outgassing. Along with H_2O and CO_2 , this process can bring trace gases to the surface, such as H_2S , SO_2 , CH_4 , NH_3 , HF , H_2 , CO and noble gases, such as Ar , Xe , etc.

On Earth and Mars, there is strong evidence that this secondary outgassing has played a major role in shaping their present atmospheres. In particular, Tian *et al.* [16] showed that the thermal escape (see below) induced by the extreme ultraviolet flux from the young sun was so strong that a CO_2 atmosphere could not have been maintained on Mars until about 4.1 billion years ago. Nevertheless, a late secondary atmosphere is thought to have been degassed, in particular via the magmas that formed the large volcanic Tharsis province. Phillips *et al.* [17] estimated that the integrated equivalent of a 1.5 bar CO_2 atmosphere could have accumulated but more realistic models have significantly lowered this value [18,19]. Similarly, 5% of photochemically unstable methane present in the present-day Titan atmosphere is thought to originate from episodic outgassing of methane stored as clathrate hydrates within an icy shell in the interior of Titan [20].

(b) Atmospheric sinks

While tens to thousands of bars of H/He and CO_2 may have been present in the early Earth atmosphere, they are obviously not there anymore (the water now being in liquid form in the oceans). This tells us that some processes, such as atmospheric escape and weathering/ingassing, can play the role of atmospheric sinks and that these processes are powerful enough to remove completely massive protoatmospheres if the right conditions are met.

Considering the fact that there are three main successive delivery mechanisms of different volatiles during the early stages of the planet's evolution, the main questions are to know when these atmospheric sinks are most efficient and if they can selectively deplete some species with respect to the others. This is what we now discuss.

(i) Atmospheric escape

Atmospheric molecules can leave the planet's attraction if they go upward with a speed exceeding the escape velocity [21]. However, in the lower part of the atmosphere, the gas density ensures a high collision rate, preventing hot particles with a sufficient velocity to leave the planet. As the density decreases with height, this assumption breaks down when the mean free path of the particles becomes bigger than the scale height of the atmosphere. Around this level, called the exobase, stellar excitation by radiation and plasma flows is important, and fast-enough atmospheric particles can actually escape.

There are several ways for the particles to reach escape velocities, defining the various escape mechanisms that can be separated into two families: thermal and non-thermal escape (see [22] for a review).

Thermal escape characterizes atmospheric escape primarily caused by the radiative excitation of the upper atmosphere. To begin with, it depends on the gravity and on the temperature of the exobase. This temperature is not controlled by the total bolometric insolation, which heats the surface and lower atmosphere, but by the flux of energetic radiation and the plasma flow from the star (especially the extreme ultraviolet, which is absorbed by the upper atmosphere). It also depends on the ability of the atmospheric molecules to radiatively cool to space by emitting infrared radiation; to simplify, greenhouse gases like CO₂ can efficiently cool, whereas other gases like N₂ cannot. Thermal escape exhibits two regimes:

- *Jeans escape*, when the exosphere is in hydrostatic equilibrium, and only the particles in the high energy tail of the Maxwell distribution can leave the planet. Lighter atoms and molecules like hydrogen and helium are more affected because they reach a much higher velocity at a given thermospheric temperature.
- *Hydrodynamic escape*, the so-called *blow-off* regime, when radiative heating can only be compensated by an adiabatic expansion and escape of the whole exosphere. On the terrestrial planets in our solar system such conditions may have been reached in H- or He-rich thermospheres heated by the strong EUV flux of the young Sun.

Non-thermal escape results from energetic chemical reactions or interactions with the stellar wind (*ion pick-up, plasma instabilities, cool ion outflow, polar wind*, etc.). See Lammer [22] for a more complete description of these processes, which can play a significant role on planets like modern Earth where gravity and temperatures prevent efficient thermal escape.

Impact escape. Finally, atmospheres can be lost to space because of the impacts of comets or asteroids. If the gravity is low enough (thus especially for small bodies), and if impactors are sufficiently big and fast, the hot plumes resulting from the impact can expand faster than the escape velocity and drive off the overlying air. On small planets and satellites, the efficiency of this process does not depend on the temperatures and the insolation, so that small bodies may not be able to keep an atmosphere even if atmospheric temperatures are very low.

While it is clearly beyond the scope of this article to go into the details of each of these mechanisms, it is interesting to derive an order of magnitude estimate for the maximum escape that can be expected for a given planet. This can be done by considering the hypothetical case of *energy limited escape*. This limit is obtained by assuming that a given fraction η (the heating efficiency) of the radiative flux available to be absorbed in the upper atmosphere is actually used to extract gas from the gravitational potential well of the planet. Because the exospheric levels are only sensitive to the very energetic photons in the X-EUV range (wavelength below 100 nm), the energy limited escape rate F_{esc} can be written as

$$F_{\text{esc}} = \eta \frac{R_p F_{\text{XUV}}}{GM_p} (\text{kg m}^{-2} \text{s}^{-1}), \quad (2.1)$$

where F_{XUV} is the averaged XUV flux received by the planet (i.e. divided by 4 compared with the flux at the substellar point), G is the universal gravitational constant, and R_p and M_p are the planetary radius and mass, respectively.

Unlike the total bolometric luminosity, the stellar luminosity in the X-EUV range F_{XUV} is correlated with the stellar activity, which is very high at young ages and declines over time. Therefore, the escape rate strongly varies with time. For example, the solar EUV flux is believed to have been 100 times stronger than that today during the first 100 million years of our Sun's life, to later decrease following a power law [8,23]. Thermal escape is thus most relevant during the first tens to hundreds of million years after the star formation, i.e. on a timescale which is similar to the atmosphere formation process! The implications of the coincidence discussed in §2c.

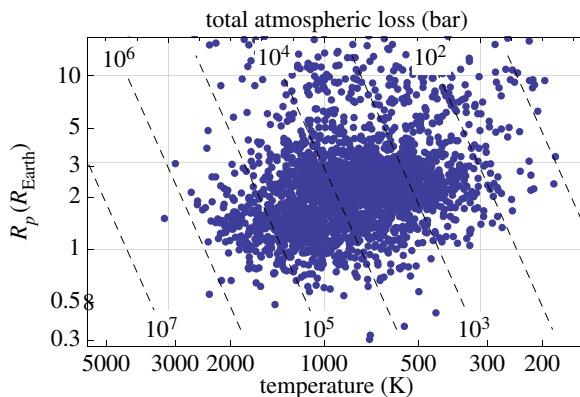


Figure 3. Kepler planet candidates in a radius-equilibrium temperature diagram. Dashed lines represent contours of the average total amount of atmosphere that could be lost by a planet at the given position in the diagram in bars (atmospheric loss is integrated over 5 Gyr and accounts for the early active phase of the star). ‘Temperature’ corresponds to the planetary equilibrium temperature, as defined in figure 1. (Online version in colour.)

To give an idea of how strong atmospheric escape can be, we computed the total integrated atmospheric pressure that can be lost during the planet lifetime,

$$p_{\text{esc}} = \int \frac{GM_p}{R_p^2} F_{\text{esc}}(t) dt = \frac{\eta}{R_p} \int F_{\text{XUV}}(t) dt, \quad (2.2)$$

as a function of the planetary radius and for an efficiency of $\eta = 0.15$ [13,22,24]. The variation over time of the XUV to bolometric flux ratio (so that the results can be expressed in terms of the equilibrium temperature of the planet) is modelled for a solar-type star using the parametrization of Sanz-Forcada *et al.* [25]. The results are shown in figure 3 (dashed labelled contours). As expected, atmospheres are more sensitive to escape when the planet receives more flux (higher equilibrium temperature) and is smaller (weaker gravity). Interestingly, current planet candidates (purple dots in figure 3) are expected to exhibit very different levels of atmospheric losses, with cold giant planets for which the effect of escape on the atmospheric content can almost be neglected and highly irradiated Earth-like objects for which the whole atmosphere has probably been blown away (see §2c; [26]). Note that we did not include in our calculations the diminution of gravity resulting from the decrease in atmospheric mass. This can induce a positive feedback which can further accelerate atmospheric loss.

In figure 3, we assume a solar-type star, but in reality, at a given bolometric insolation, escape is also expected to be more intense around low-mass stars as they emit a larger fraction of their flux in the XUV range. This can be seen in figure 4. It is because of the increased duration of the active phase of lower mass stars.

(ii) Weathering and ingassing

The atmospheric composition can also be altered by interactions with the surface. Indeed, if the right conditions are met, some constituents of the atmosphere can chemically react with the surface and get trapped there. In addition to this process, called weathering, the aforementioned surface can be buried by lava flows or subducted by plate tectonics, re-enriching the mantle in the volatiles that have been trapped; volatiles that will eventually be released by volcanic activity later.

An example of such a process is the chemical weathering of silicate minerals in the presence of liquid water and CO_2 . Atmospheric CO_2 goes into solution in liquid water relatively easily and the resulting fluid, carbonic acid (H_2CO_3), reacts with the silicate minerals of the crust to weather

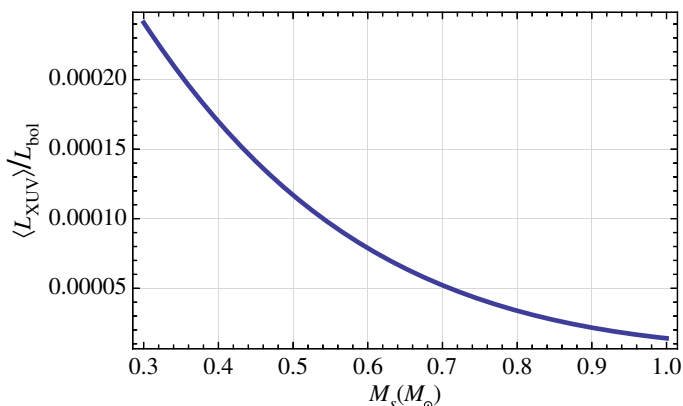
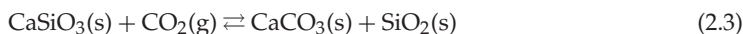


Figure 4. Ratio between the mean XUV luminosity L_{XUV} (between 0.1 and 100 nm and integrated over the first 5 Gyr of the star) and the bolometric luminosity L_{bol} of a star as a function of its mass. This is computed using Sanz-Forcada *et al.* [25] parametrization for L_{XUV} . The greater ratio for lower mass stars stems from the longer early activity phase and results in a more efficient escape around smaller stars (at a given bolometric flux). (Online version in colour.)

the rock and release calcium, magnesium and iron ions into the water. These ions can promote the precipitation of solid carbonates, $(Ca,Mg,Fe)CO_3$. To simplify, the following net reaction



can occur. This reaction traps carbon dioxide into carbonates that can accumulate before being buried by subduction. A very interesting property of this *carbonate–silicate cycle* is that it provides a powerful stabilizing feedback on planetary climates on geological timescales [27], as detailed in §5*b*. On the Earth, most of the initial CO_2 inventory is thought to be trapped in the form of carbonates in the crust after chemical precipitation. The formation of carbonates has also been suggested on early Mars when abundant liquid water seems to have flowed on the surface, possibly explaining the fate of an early thick CO_2 atmosphere [28]. However, almost no carbonates were initially detected by the OMEGA imaging spectrometer in spite of its high sensitivity to the spectral signature of carbonates [29]. Recently, several observations from orbiters [30,31] and landers [32,33] have revived the carbonate hypothesis and reasserted the importance of carbon dioxide chemistry in martian climate history [34].

(c) Major classes of atmospheres

Let us make an attempt to see how the processes described above fit together to produce the diversity of atmospheres that we know or can expect. Indeed, we have seen that they have the ability to build or get rid of an entire atmosphere. The key is now to identify the mechanism(s) that are most relevant to a given planetary environment. The result of this process is illustrated in figure 5 and discussed below.

(i) H/He dominated

Hydrogen and helium being the lightest elements and the first to be accreted, they can most easily escape. The occurrence of H/He-dominated atmospheres should thus be limited to objects more massive than the Earth. Indeed, in the Solar System, none of the terrestrial planetary body managed to accrete or keep a potential primordial H/He envelope, even the coldest ones which are less prone to escape.

Figure 2 suggests that a mass as low as $2M_{Earth}$ can be sufficient to build and keep such an atmosphere. But it also suggests that being more massive than that is by no means a sufficient condition. Indeed, some objects have a bulk density similar to the Earth up to $8\text{--}10M_{Earth}$. Although these high-density planets receive a stronger stellar insolation on average, it is not clear

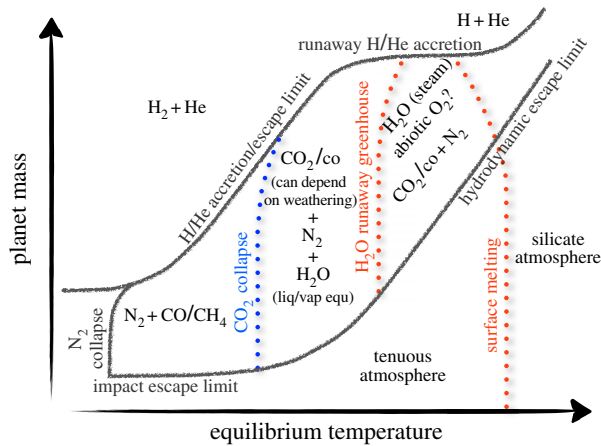


Figure 5. Schematic summary of the various class of atmospheres. Each line represent a transition from one regime to another, but note that these ‘transitions’ are in no way hard limits. Only the expected dominant species are indicated, but other trace gas should of course be present. (Online version in colour.)

yet whether this correlation stems from the fact that planets forming in closer orbits can accrete less nebular gas [7] or from the fact that hotter planets exhibit higher escape rates. Note that the first hypothesis assumes that such close-in planets are formed *in situ*.

Then, the presence of a large fraction of primordial nebular gas in the atmosphere of warm to cold planets above a few Earth masses should be fairly common. The real question will be to know the atmosphere mass and by how much these atmospheres will be enriched in heavy elements compared with the parent star. Such information will be critical to better understand the early stages of planet and atmosphere formation during the nebular phase.

(ii) $\text{H}_2\text{O}/\text{CO}_2/\text{N}_2$ atmospheres

Then, if, for some reason, the planet ends up its accretion phase with a thin enough H/He atmosphere so that surface temperatures can be cold enough for the solidification of the rocky surface, a significant amount of H_2O and CO_2 should be released (envelope between grey curves in figure 5). To understand what will happen to these volatiles, however, one needs to understand in which climate regime the planet will settle.

Saving for later the very hot temperatures for which the surface itself is molten, let us go through the different available regimes from the upper left to the lower right part of figure 5.

Above a certain critical flux, the so-called runaway greenhouse limit, the positive radiative feedback of water is so strong that the atmosphere warms up until surface water is vapourized [35].¹ In this case, the absence of surface water hampers CO_2 weathering, leaving most of the CO_2 inventory in the atmosphere.

In this case, a key question concerns the conservation of the water itself. Indeed, if H_2O is a major constituent of the atmosphere, it can easily be photo-dissociated high up. This produces H atoms that are ready to escape. Although this seemed to occur on Venus, more massive planets with a higher gravity to counteract escape or objects that accreted more water may still possess a significant fraction of atmospheric water (see, for example the debate on the atmospheric composition of GJ1214b; [38,39])

Below the runaway greenhouse limit, water can condense at the surface. Except for a few planets very near the limit, water vapour should thus remain a trace gas in liquid/vapour equilibrium with the surface. Thus, the atmosphere could be dominated by species that are less abundant in the initial inventory but have been slowly outgassed, namely N_2 , among others. In such a state, CO_2 weathering can be efficient so that the amount of CO_2 might depend

¹Although this limit is not that well defined when the water inventory is very limited [36,37].

on the surface temperature (see §5*b*). However, if water is lost owing to atmospheric escape, especially for lower mass planets, for example, as Mars, or hotter ones, CO₂ could build up in the atmosphere and become the dominant gas.

For colder climates, even CO₂ greenhouse warming is insufficient to prevent its condensation indefinitely. When this CO₂ collapse occurs, water, of course, but also CO₂ itself can only be found in trace amounts. As is seen in the solar system (e.g. Titan), N₂ thus becomes the only stable abundant species (apart from H₂/He). Carbon compounds can be found in the form of CO or CH₄ depending on the oxidizing/reducing power of the atmosphere. This can continue until the triple point of nitrogen itself is reached. At that point, N₂ ice albedo feedback favours a very cold climate where nitrogen is in condensation/sublimation equilibrium with the surface, leaving only thin atmospheres, such as the ones found on Pluto and Triton.

(iii) Photochemistry and CO

CO₂ atmosphere may not be photochemically stable. In fact, the abundance of CO₂ on Mars and Venus seemed puzzling early in the space age [40] because CO₂ is readily photo-dissociated. The direct three-body recombination, $\text{CO} + \text{O} + \text{M} \rightarrow \text{CO}_2 + \text{M}$, is spin-forbidden, and therefore extremely slow at atmospheric temperatures. The solution for Mars is that photolysis of water vapour produces OH radicals that react readily with CO to make CO₂; in effect, water vapour photolysis catalyses the recombination of CO₂. Could the equilibrium be reversed in favour of CO in some conditions? Zahnle *et al.* [41] showed that this may happen in thick cold (and thus relatively dry) atmosphere, although they noted that in reality CO could react with the surface (Fe) and be recycled as CO₂ by another path. Another interesting point is that the stability question is asymmetric. Under plausible conditions, a significant CO atmosphere can be converted to a CO₂ atmosphere quickly in the case of any event (impact, volcanism) that may provide water vapour, whereas it takes tens to hundreds of million years to convert from CO₂ back to CO. In any case, the behaviour of CO atmosphere would be somewhat different from that of CO₂. CO is a weaker greenhouse gas than CO₂, but it condenses at a significantly lower temperature (or higher pressure) than CO₂. In very cold cases, conversion to CO may thus prevent atmospheric collapse into CO₂ ice glaciers.

(iv) The possibility of abiotic O₂

Molecular oxygen O₂ cannot easily become a dominating species in a planetary atmosphere because it is chemically reactive and is not among the volatile species provided by planetesimals. Most of the O₂ in Earth's present atmosphere is thought to have been produced by biological oxygenic photosynthesis. Nevertheless, several abiotic scenarios that could lead to oxygen-rich atmospheres have been suggested and studied in detail, because the presence of O₂ and the related species O₃ (easier to detect) in the atmosphere of an exoplanet are considered to be possible biomarker compounds [42,43].

The most likely situation in which O₂ might accumulate and become a dominant species is a runaway greenhouse planet, like early Venus, on which large amounts of hydrogen escape from a hot, moist atmosphere (see [8] and references therein). Because the hydrogen originates from H₂O, oxygen is left behind. The escape of a terrestrial ocean equivalent of hydrogen, unaccompanied by oxygen sinks, could leave an atmosphere containing up to 240 bars of O₂ [44]. Alternatively, O₂ could be produced by photolysis of CO₂ in a very dry environment, but its concentration is then not likely to reach more than a few per cent [44,45].

(v) Thin silicate atmospheres

For low-mass objects and very hot objects (lower part of figure 5), escape is supposed to be efficient. Bodies in this part of the diagram are thus expected to have no atmosphere or possibly a very tenuous exosphere. This class actually encompasses many Solar System bodies: Mercury, the Moon, Ganymede, Callisto, etc. On such bodies, a tenuous gaseous envelope can be maintained

by the release of light molecules and atoms from the surface because of the energetic radiation and charged particles impacting the surface (e.g. O₂ on Ganymede and Europe) or the release of gases, such as radon and helium, resulting from radiative decay within the crust and mantle (e.g. Argon and Helium on the Moon). These are not atmospheres.

An interesting case is Io, which is characterized by an intense volcanic activity resulting from tidal heating from friction generated within Io's interior as it is pulled between Jupiter and the other Galilean satellites. This activity allows for the formation of an extremely thin and varying atmosphere consisting mainly of sulfur dioxide (SO₂).

In extrasolar systems, another exotic situation can arise. Indeed, some planets, such as CoRoT-7b [46], are so close to their host star that the temperatures reached on the dayside are sufficient to melt the rocky surface itself. As a result, some elements usually referred to as 'refractory' become more volatile and can form a thin 'silicate' atmosphere [26,47]. Depending on the composition of the crust, the most abundant species should be, by decreasing abundance, Na, K, O₂, O and SiO. Interestingly enough, the energy-redistribution effect of such an atmosphere could be limited to the day side of the planet as condensation occurs rapidly near the terminators [48]. In addition, silicate clouds could form. Both of these effects should have a significant impact on the shape of both primary- and secondary-transit lightcurves, allowing us to constrain this scenario in the near future.

3. On the importance of planetary rotation

(a) Rotation and climate

Besides the atmospheric composition and the mean insolation, one of the key parameters that determines a planetary climate is the rotation of the body (period, obliquity). Rotation rate and obliquity are thought to influence the climate in two ways. On the one hand, they govern the latitudinal distribution of insolation as well as the seasonal and diurnal cycle. On the other hand, modelling studies, laboratory experiments and our experience in the solar system show that the atmospheric circulation and transport directly depends on the rotation rate via the Coriolis and centrifugal forces. They control the extension of the Hadley circulation and the formation of extratropical jets (in the strongly rotating regime), the type of planetary waves, and the tendency of slowly rotating planets towards super-rotation (see [49,50] and references therein).

Because of the angular momentum accreted during their formation, most planets initially tend to rotate around their axis relatively quickly. In the solar system, all planets that have not been significantly influenced by the gravitation of another body rotate with a period of about one Earth day or less (e.g. Mars, Jupiter, Saturn, Uranus, Neptune). However, during its existence, the rotation of a body is modified by tidal effects resulting from gravitational forces from its parent star or from its satellites (or from its parent planet in the cases of satellites). These forces tend to cancel out the obliquity (creating poles where almost no starlight reaches the surface) and synchronize the rotation rate (possibly creating a permanent night side). They can thus strongly influence the climate.

(b) Tidal evolution of planetary spin

In which cases will a planet be affected by gravitational tides? When an extended, deformable body is orbited by another mass, the differential gravitational attraction of the latter always causes the primary object to be distorted. These periodic deformations create friction inside the deformable body, which dissipates mechanical energy and allows angular momentum to be exchanged between the orbit and the spin of the two orbiting objects. In general, such tidal interactions eventually lead to an equilibrium state where the orbit is circular and the two components of the system are in a spin-orbit synchronized state with a zero obliquity [51]. However, because the tidal potential decreases in proportion to the distance between the two bodies to the minus sixth power, this equilibrium can take several dozens of billions of years to be achieved.

In a star–planet system, because the angular momentum contained in the planetary spin is small compared with the orbital and stellar one, the evolution of the planetary spin (synchronization and alignment) is the most effective and occurs first. Indeed, the standard theory of equilibrium tides [52–54] predicts that a planet should synchronize on a timescale equal to

$$\tau_{\text{syn}} = \frac{1}{3} r_g^2 \frac{M_p a^6}{GM_s^2 R_p^3 k_{2,p} \Delta t}, \quad (3.1)$$

where M_p and R_p are the planetary mass and radius, respectively, M_s is the stellar mass, r_g is the dimensionless gyration radius ($r_g^2 = 2/5$ for a homogeneous interior; [55]), G is the universal gravitational constant, $k_{2,p}$ the tidal Love number of degree 2 that characterizes the elastic response of the planet and Δt a time lag that characterizes the efficiency of the tidal dissipation into the planet's interior (the higher Δt , the higher the dissipation). While the exact magnitude of the tidal dissipation in terrestrial planets remains difficult to assess, one can have a rough idea of the orders of magnitude involved by using the time lag derived for the Earth from the analysis of Lunar Laser Ranging experiments, $k_{2,p} \Delta t = 0.305 \times 629$ s [56]. This yields a synchronization timescale of 20 Gyr for the Earth, 3 Gyr for Venus and 80 Myr for Mercury. This is consistent with the fact that, while the Earth has been able to keep a significant obliquity and a rapid rotation, both Venus and Mercury have a rotation axis aligned with the orbit axis and a slow rotation, although this rotation is not synchronous (see below). The fact that Mercury's orbit is still eccentric confirms that tidal circularization proceeds on a longer timescale. In many extrasolar systems where planets are found much closer to the central star, tidal synchronization and coplanarization of the planetary rotation is thus expected to be fairly common. In particular, rocky planets within or closer than the habitable zone of M and K stars are thought to be significantly tidally evolved. This can have a profound impact on the climate of these planets, as it creates permanent cold traps for volatiles at least near the poles, and possibly on the permanent dark side if the rotation rate is fully synchronized.

However, when a terrestrial planet has a permanent bulk mass distribution asymmetry or possesses a thick atmosphere, tidal synchronization is not the only possible spin state attainable by the planet. In the first case, as for Mercury, if the planet started from an initially rapidly rotating state, it could become trapped in multiple spin orbit resonances during its quick tidal spin down because of its eccentric orbit. This is also expected for extrasolar planets. Because Gl581d has a non-negligible eccentricity, it has a high probability of being captured in 3:2 (three rotation per two orbits) or higher resonance before reaching full synchronization [57]. In the second case, thermal tides in the thick atmosphere can create a torque that drives the planet out of the usual tidal equilibrium. This is what is thought to cause the slow retrograde rotation of Venus [58,59], and it could also even lead extrasolar terrestrial planets out of synchronization [60]. In any cases, these states all have a low obliquity, which would have an important impact on the climate by creating cold poles.

4. Which climate for a given atmosphere?

(a) Key processes in a climate system

Any planetary atmosphere exhibits an apparent high level of complexity, owing to a large number of degrees of freedom, the interaction of various scales, and the fact that atmospheres tend to propagate many kind of waves.

However, the key physical and dynamical processes at work on a terrestrial planet are in finite number. To begin with, on most planets, the following coupled dynamical and physical processes control the climate (figure 6):

- (1) Radiative transfer of stellar and thermal radiation through gas and aerosols.
- (2) The general circulation of the atmosphere primarily forced by the large-scale, radiatively induced temperature gradients.

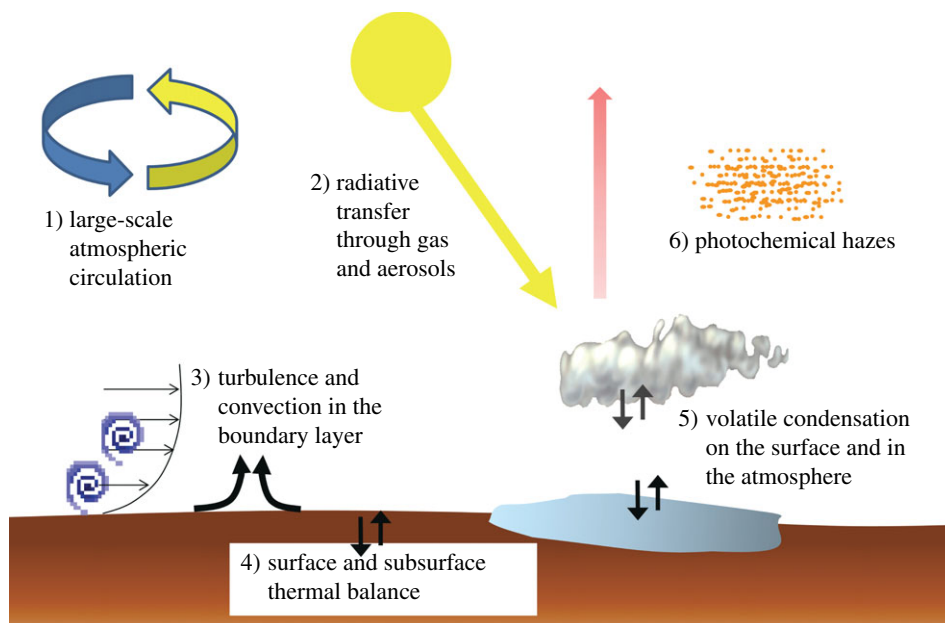


Figure 6. The key physical and dynamical processes which control a terrestrial planet climate. In practice, when modelling planetary climates, these processes can be parametrized independently and combined to create a realistic planetary global climate model. (Online version in colour.)

- (3) Vertical mixing and transport owing to small-scale turbulence and convection.
- (4) The storage and conduction of heat in the subsurface.
- (5) The phase changes of volatiles on the surface and in the atmosphere (clouds and aerosols).
- (6) To this list, one could add a catalogue of processes which are only relevant in particular cases, or which play a secondary role: photochemistry (producing aerosols, hazes or creating spatial inhomogeneities in the atmospheric composition), mineral dust lifting, oceanic transport, molecular diffusion and conduction (at very low pressure), etc.

Depending upon the planet's physical characteristics (orbit, size, rotation, host star, etc.) and of the composition of its atmosphere, the combination of all these processes can lead to a variety of climates that we will not try to describe here. Instead, we discuss below our ability to simulate and predict the diversity of these terrestrial climates, using numerical models.

(b) Modelling terrestrial planetary climate

(i) From one- to three-dimensional realistic models

The processes listed above can be described with a limited number of coupled differential equations, and it is now possible to develop numerical climate models to predict the environment on terrestrial planets.

Until recently, a majority of studies on terrestrial exoplanets had been performed with simple one-dimensional steady-state radiative convective models. They can evaluate the global mean conditions on a given planet resulting from the radiative properties of its atmosphere and the insolation from its star (see for instance the reference paper on habitability by Kasting *et al.* [61]). Such one-dimensional models have been extremely useful to explore the possible climate regimes, although they are often not sufficient to predict the actual state of a planet, and in particular represent the formation, distribution and radiative impact of clouds or to simulate

local conditions at a given time (for instance, owing to the diurnal and seasonal cycles). Three-dimensional models are especially necessary to estimate the poleward and/or night side transport of energy by the atmosphere and, in principle, the oceans.

Exploring and understanding the atmospheric transport and the possible circulation regime as a function of the planet characteristics is a research field by itself. It does not require complete realistic climate models. For this purpose, dynamicists have used models with a three-dimensional hydrodynamical ‘core’ designed to solve the Navier–Stokes fluid dynamical equations in the case of a rotating spherical envelope, forced with simplified physics to represent the possible thermal gradients (see a review in [49] as well as the recent work in earlier studies [50,62,63]).

More complete three-dimensional numerical global climate models (‘GCMs’) can be built by combining the various components which are necessary to simulate the major processes listed above (figure 6): a three-dimensional hydrodynamical ‘core’, and, for each grid-point of the model, a radiative transfer solver, a parametrization of the turbulence and convection, a subsurface thermal model, a cloud model, etc. Such models have been developed (in some cases for more than 20 years) for the telluric atmospheres in the solar system: the Earth (of course), Mars Venus, Titan, Triton and Pluto. The ambition behind the development of these GCMs is high: the ultimate goal is to build numerical simulators only based on universal physical or chemical equations, yet able to reproduce or predict all the available observations on a given planet, without any ad hoc forcing. In other words, we aim at creating in our computers virtual planets ‘behaving’ exactly like the actual planets. In reality of course, nature is always more complex than expected, but one can learn a lot in the process. In particular, a key question is now to assess whether the GCM approach, tested in the solar system, is ‘universal’ enough to simulate the diversity of possible climate on terrestrial exoplanets and accurate enough to predict the possible climate in specific cases.

Several teams are now working on the development of three-dimensional global climate models designed to simulate any type of terrestrial climate, i.e. with any atmospheric cocktail of gases, clouds and aerosols, for any planetary size and around any star. For instance, at LMD, we have recently developed such a tool (e.g. [64–67]), by combining the necessary parametrizations listed above. One challenge has been to develop a radiative transfer code fast enough for three-dimensional simulations and versatile enough to model any atmospheric composition accurately. For this purpose, we used the correlated-k distribution technique. We also included a dynamical representation of heat transport and sea-ice formation on a potential ocean, from [68].

(ii) What have we learned from our experience in the solar system?

The first lesson from the modelling of the (limited) diversity of climate in the solar system is that the same equations are often valid in several environments and that the different model components that make a climate model can be applied without major changes to most terrestrial planets. Of course, the spectroscopic properties of the atmosphere, for instance, must be adapted in each case. Atmospheric radiative properties have been well studied in the Earth’s case for which numerous spectroscopic databases are available. However, some unknowns remain for observed atmospheres like Mars or Venus, and many more uncertainties affect the modelling of theoretical exotic atmospheres not yet observed (e.g. hot and wet atmospheres with a high partial pressure of water vapour). The parametrization of large-scale dynamics, turbulent mixing and subsurface heat conduction have been applied to different planets without modification and comparisons with the available observations have not revealed major problems. In some cases, some simplifications that were initially done for the Earth’s case (constant atmospheric composition, ‘thin atmosphere approximation’) must be questioned on other planets. For instance, on Venus the air-specific heat varies significantly (around 40%) with temperature from the surface to the atmosphere above the clouds, whereas it was assumed to be constant in the dynamical cores derived from the Earth modelling. The consequences on the potential temperature and dynamical core were discussed in [69].

The second major lesson is that, by many measures, global climate models work. They have been able to predict the behaviour of many aspects of several climate systems on the basis of physical equations only. Listing the success of global climate models on other planets is out of the scope of this paper, but we could mention a few examples (with a bias towards our models developed at LMD). On Mars, assuming the right amount of dust in the atmosphere it has been relatively easy to simulate the thermal structure of the atmosphere and the behaviour of atmospheric waves, such as thermal tides and baroclinic waves [70–76], to reproduce the main seasonal characteristics of the water cycle [77,78] or to predict the detailed behaviour of ozone [79,80]. On Titan, GCMs have anticipated the super-rotating wind fields with amplitude and characteristics comparable to observations [81,82] and allowed to simulate and interpret the detached haze layers [83,84], the abundance and vertical profiles of most chemical compounds in the stratosphere and their enrichment in the winter polar region [85], the distribution of clouds [86] or the detailed thermal structure observed by Huygens in the lowest 5 km [87]. On Venus, the development of ‘full’ GCMs (i.e. at least coupling a three-dimensional dynamical core and a realistic radiative transfer) is more recent, but these models successfully reproduce the main features of the thermal structure and the super-rotation of the atmosphere [69,88–90].

The third and even more interesting lesson is related to the ‘failure’ of planetary global climate models [66]. When and why GCMs have not been able to predict the observations accurately? Different sources of errors and challenges are listed below.

- *Missing physical processes.* As can be expected, in many cases GCMs fail to accurately simulate an observed phenomenon simply because a physical process is not included in the GCM. For instance, for many years, the thin water ice clouds present in the Martian atmosphere had been assumed to have a limited impact on the Martian climate. Recently, several teams have included their effect in GCM simulations [91–94].
What they found is that not only do the clouds affect the thermal structure locally, but that their radiative effects could solve several long-lasting Mars climate enigmas, such as the pause in baroclinic waves around winter the solstice, the intensity of regional dust storms in the northern mid-latitudes or the strength of a thermal inversion observed above the southern winter pole.
- *Positive feedbacks and instability.* Another challenge is present for climate modellers when the system is very sensitive to a parameter because of positive feedbacks. A well-known example is the albedo of snow and sea-ice on the Earth. If one tries to model the Earth climate systems from scratch, it is rapidly obvious that this model parameter must be tuned to ensure a realistic climate at high and mid-latitude. An overestimation of the ice albedo results in colder temperatures, more ice and snow, etc.
- *Nonlinear behaviour and threshold effects.* An extreme version of the model sensitivity problem is present when the climate depends on processes which are nonlinear or which depend on poorly understood physics. For instance, the main source of variability in the Mars climate system is related to the local, regional and sometimes global dust storms that occur on Mars in seasons and locations that vary from year to year. This dust cycle remains poorly understood, possibly because the lifting of dust occurs above a local given wind threshold stress which may or may not be reached depending on the meteorological conditions. As a result, modelling the dust cycle and in particular the interannual variability of global dust storms remains one of the major challenges in planetary climatology—not mentioning a hypothetical ability to predict the dust storms [95–97]. Most likely, in addition to the threshold effect, a physical process related to the evolution of the surface dust reservoirs is missing in the models.
- *Complex sub-grid scale processes.* Another variant of the problems mentioned above can be directly attributed to processes which cannot be resolved by the dynamical core but which play a major role in the planetary climate. Mars dust storms would be once again a good example, but the most striking example is the representation of sub-grid scale clouds in the Earth GCMs. The parametrization of clouds has been identified as a major

source of disagreement between models and uncertainties when predicting the future of our planet (e.g. [98]).

- *Weak forcings, long timescales.* While different GCMs can easily agree between themselves and with the observations when modelling a system strongly forced by the variations of, say, insolation, GCM simulations naturally become model sensitive when the evolution of the system primarily depends on a subtle balance between modelled processes.

An interesting case is the Venus general circulation—the super-rotation of Venus atmosphere is the result of a subtle equilibrium involving balance in the exchanges of angular momentum between surface and atmosphere, and balance in the angular momentum transport between the mean meridional circulation and the planetary waves, thermal tides and gravity waves. Comparative studies between Venus GCMs under identical physical forcings [99,100] have recently shown that modelling this balance is extremely sensitive to the dynamical core details, to the boundary conditions and possibly also to initial conditions. These studies revealed that various dynamical cores, which would give very similar results in Earth or Mars conditions, can predict very different circulation patterns in Venus-like conditions.

- *The take-home message.* Overall, our experience in the Solar System has shown that the different model components that make a climate model can be applied without major changes to most terrestrial planets. It has also revealed potential weaknesses and inaccuracies of GCMs. Clearly, when modelling climate systems which are poorly observed, it is necessary to carefully explore the sensitivity of the modelled system to key parameters, in order to ‘bracket’ the reality. Nevertheless, it seems to us that when speculating about the climate regime on a specific detected planet and in particular its habitability, the primary uncertainty lies in our ability to predict and imagine the possible nature of its atmosphere, rather than in our capacity to model the climate processes for a given atmosphere. This said, that does not mean that it is easy to predict temperatures and the states of the volatiles, because in many cases an accurate climate model exhibits a very high sensitivity to some parameters, as detailed in §5.

5. Climate instability and feedbacks

(a) Runaway glaciation and runaway greenhouse effects

While studying the sensitivity of Earth’s climate and the extension of the habitable zone (i.e. the range of orbits where the climate can be suitable for surface liquid water and life), it has been discovered that the climate of a planet with liquid water on its surface can be extremely sensitive to parameters such as the radiative flux received from its parent star. This results from the fact that the radiative effect of water strongly varies with temperature as its phase changes, inducing strong feedbacks [61].

For instance, slightly ‘moving’ a planet like the Earth away from the Sun induces a strong climate instability because of the process of ‘runaway glaciation’: a lower solar flux decreases the surface temperatures, and thus increases the snow and ice cover, leading to higher surface albedos which tend to further decrease the surface temperature [101–103]. The Earth would become completely frozen (and several tens of degrees colder on average) if moved away from the Sun beyond a threshold distance which is highly model-dependent, but probably close to the present orbit (5–15% further from the Sun). Furthermore, there is an hysteresis. This Earth-like planet would remain ice-covered when set back to its initial conditions: being completely frozen is thus, in theory, another stable regime for the Earth. Note that this ice-snow albedo feedback is much weaker around M-stars because water ice tends then to have a much lower albedo, since it absorbs in the near-infrared where M-stars emit a significant fraction of their radiation [104].

Alternatively, when a planet with liquid water on its surface is ‘moved’ towards its sun, its surface warms, increasing the amount of water vapour in the atmosphere. This water vapour strongly enhances the greenhouse effect, which tends to further warm the surface. This ‘runaway greenhouse’ process can destabilize the climate on the basis of simple one-dimensional model calculations, Kasting [35] found that on an Earth-like planet around the Sun, oceans would completely vapourize below a threshold around 0.84 Astronomical Units (AU). He also showed that the stratosphere would become completely saturated by water vapour at only 0.95 AU. There, it could be rapidly dissociated by ultraviolet radiation, with the hydrogen lost to space (the Earth currently keeps its water thanks to the cold-trapping of water at the tropopause).

In all these examples, a lot of uncertainties exist, especially in relation to the role of clouds, the actual ice-snow albedo, the spectroscopy of water vapour, the transport of heat by the atmosphere and ocean, etc [37,105,106]. The threshold values obtained are highly model-dependent. Nevertheless, these famous cases tell us that real climates systems can be affected by strong instabilities which can drive planets subjected to a similar volatiles inventory and forcing to completely different states. This is probably not limited to water. At colder temperatures, the concept of runaway greenhouse and runaway glaciation can be extended to CO₂ (which can influence the albedo or the atmospheric greenhouse effect by condensing onto the surface or subliming), or even N₂ [107].

The uncertainties related to the volatiles phase changes are even higher in the cases of tidally evolved planets (see §3), like in the habitable zone around M-stars. Then the permanent night side (for a planet locked in a 1 : 1 resonance) or at least the permanently cold polar regions (for planet with very small obliquities) are cold traps on which water and other condensable atmospheric gases like CO₂ or N₂, may permanently freeze, possibly inducing an atmospheric collapse. For a given planet, this allows, in theory, additional stable climate solutions (e.g. [67,108,109]). Nevertheless, it can be noted that on such slowly rotating bodies the atmospheric dynamics can be very efficient to transport heat from the sunlit regions to the night side and therefore homogenize the temperatures and prevent atmospheric collapse [64,108,110]. In fact, the habitability of planets around an M-dwarf could actually ‘benefit’ from tidal locking. The cold trap on the night side may allow some water to subsist well inside the inner edge of the classical habitable zone. If a thick icecap can accumulate there, gravity-driven ice flows and geothermal flux could come into play to produce long-lived liquid water at the edge and/or bottom of the icecap [67]. Similarly, on a cold-locked ocean planet, the temperature at the substellar point can be much higher than the planetary average temperature. An open liquid water pool may form around the substellar point within an otherwise frozen planet [109].

(b) Climate stabilization and plate tectonics

Another concept that must be taken into account when speculating about the possible climates on terrestrial planets is the possibility that a planet can be influenced by negative feedbacks, which will ultimately control its atmosphere and drive it into a specific regime.

For instance, such a scenario is necessary to explain the long-term habitability of the Earth, which has been able to maintain liquid on its surface throughout much of its existence in spite of a varying solar luminosity and changes in its atmospheric composition which could have led it to runaway glaciation.

Most probably, this has been possible thanks to a long-term stabilization of the surface temperature and CO₂ level owing to the carbonate–silicate cycle [61,111]. As mentioned in §2*b*(ii), on Earth, CO₂ is permanently removed from the atmosphere by the weathering of calcium and magnesium silicates in rocks and soil, releasing various ions, including carbon ions (HCO₃⁻, CO₃²⁻). These ions are transported into the world ocean through river or ground water runoff. There, they form carbonate and precipitate to the seafloor to make carbonate sediments. Ultimately, the seafloor is subducted into the mantle, where silicates are reformed and CO₂ is released and vented back to the atmosphere by volcanos. Assuming that weathering is an increasing function of the mean surface temperatures (through a presumed enhanced role of

the water cycle, precipitation, run-off, with higher temperatures) one can see that this cycle can stabilize the climate, because the abundance, and thus the greenhouse effect of CO₂ increases with decreasing temperatures, and vice versa. This mechanism is thought to be efficient for any sea–land fraction, although the climate stabilization may be limited on pure waterworld without subaerial land on which temperature-dependent weathering may occur [112].

On the Earth, the key process allowing the carbonate–silicate cycle—and more generally the long-term recycling of atmospheric components chemically trapped at the surface—is thus plate tectonics. This is a very peculiar regime induced by the convection in the mantle, the failure of the lithosphere (the ‘rigid layer’ forming the plates that include the crust and the uppermost mantle) and surface cooling.

How likely is the existence of plate tectonics elsewhere? In the solar system, Earth plate tectonics is unique and its origin is not well understood. Other terrestrial planets or satellites are characterized by a single ‘rigid lid’ plate surrounding the planet, and this may be the default regime on extrasolar terrestrial planets. On planets smaller than the Earth (e.g. Mars), the rapid interior cooling corresponds to a weak convection stress and a thick lithosphere, and no plate tectonics is expected to be maintained in the long term. On larger planets (i.e. ‘Super-Earths’), available studies have reached very different views [113–116]. It is possible that in super-Earth the large planetary radius acts to decrease the ratio of convective stresses to lithospheric resistance [115] and that the very high internal pressure increases the viscosity near the core–mantle boundary, resulting in a highly ‘sluggish’ convection regime in the lower mantles of those planets, which may reduce the ability of plate tectonics [117,118].

What these studies highlight is the possibility that the Earth may be very ‘lucky’ to be in an exact size range (within a few per cent) that allows for plate tectonics. Furthermore, Venus, which is about the size of the Earth but does not exhibit plate tectonics, shows that the Earth case may be rare and that many factors control the phenomenon. On Venus, for instance, it is thought that the mantle is drier than on Earth and that consequently it is more viscous and the lithosphere thicker [117,119]. Similar considerations led [120] to conclude that the likelihood of plate tectonics is also controlled largely by the presence of surface water. Plate tectonics may also strongly depend on the history and the evolution of the planet. Using their state-of-the-art model of coupled mantle convection and planetary tectonics, Lenardic & Crowley [121] found that multiple tectonic modes could exist for equivalent planetary parameter values, depending on the specific geologic and climatic history.

6. Conclusion

Based on the examples observed in the solar system, and on the available observations of exoplanets, we can expect a huge diversity among exoplanetary climates. In the absence of direct observations, one can only speculate on the various possible cases. According to models, climate should primarily depend on (i) the atmospheric composition and mass and the volatiles inventory (including water); (ii) the incident stellar flux (i.e. the distance to the parent star); and (iii) the tidal evolution of the planetary spin (which also depends on the distance to the star for a planet).

In theory, the atmospheric composition and mass depends on complex processes which are difficult to model: origins of volatiles, atmospheric escape, geochemistry, long-term photochemistry, etc. Some physical constraints exist, which can help us to speculate on what may or may not exist, depending on the planet size, its final distance from its star and the star type and its activity (figure 5). Nevertheless, the diversity of atmospheric composition remains a field for which new observations are necessary. Once a type of atmosphere can be assumed, theoretical three-dimensional climate studies, which benefit from our experience in modelling terrestrial atmospheres in the solar system, should allow us to estimate the range of possibilities, and in particular estimate whether liquid water can be stable on the surface of these bodies and whether they can be habitable. Whatever the accuracy of the models, predicting the actual climate regime on a specific planet will remain challenging because climate systems are

affected by strong positive feedbacks (runaway glaciation and the runaway greenhouse effect) which can drive planets subjected to a similar volatiles inventory and forcing and to completely different states.

We can hope that in the future it will be possible to learn more about exoplanetary atmospheres thanks to telescopic observations and spectroscopy. An important step will be achieved in the next decade by space telescopes like the James Webb Space Telescope or the proposed ECHO mission [122], as well as by the Earth-based telescopic observations using new generation telescope like the European Extremely Large Telescope. These projects will notably be able to perform atmospheric spectroscopy on exoplanets transiting in front of their star as seen from the Earth. Characterizing atmospheres of terrestrial planets in or near the habitable zone will remain challenging. Furthermore, the number of observable planets at a suitable distance will probably be very low. Nevertheless, well before the time when we will be able to detect and characterize a truly habitable planet, the first observations of terrestrial exoplanet atmospheres, whatever they show, will allow us to make a major progress in our understanding of planetary climates, and therefore in our estimation of the likelihood of life elsewhere in the universe.

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