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# Exoplanet interiors and habitability

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## ABSTRACT

More than 1000 exoplanets with a radius smaller than twice that of the Earth are currently known, mainly thanks to space missions dedicated to the search of exoplanets. Mass and radius estimates, which are only available for a fraction (~ 10%) of the exoplanets, provide an indication of the bulk composition and interior structure and show that the diversity in exoplanets is far greater than in the Solar System. Geophysical studies of the interior of exoplanets are key to understanding their formation and evolution, and are also crucial for assessing their potential habitability since interior processes play an essential role in creating and maintaining conditions for water to exist at the surface or in subsurface layers. For lack of detailed observations, investigations of the interior of exoplanets are guided by the more refined knowledge already acquired about the Solar System planets and moons, and are heavily based on theoretical modelling and on studies of the behaviour of materials under the high pressure and temperature conditions in planets. Here we review the physical principles and methods used in modelling the interior and evolution of exoplanets with a rock or water/ice surface layer and identify possible habitats in or on exoplanets.

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Exoplanets; interiors; evolution; habitability



## 1. Introduction

With the advent of space missions dedicated to the search of planets outside our Solar System, such as the NASA Kepler mission launched in 2009 [1,2], the number of exoplanets detected has sharply risen. About 3800 planets are now firmly identified and several thousand more need further confirmation (see [exoplanet.eu](http://exoplanet.eu) for a database of exoplanets [3]). Survey studies indicate that planets are very common around stars. The

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probability of finding an exoplanet is almost 100% for main sequence stars with masses in the range of 0.5 to 1.2 solar masses, and a significant although not well determined fraction of stars has multiple planets [4,5]. Although the exploration of the exoplanetary world only recently started, it can already confidently be concluded that there are more planets in the Milky Way than stars. Could there be life on any of these planets? We currently have no observational indication of extraterrestrial life, but ‘absence of evidence is not evidence of absence’ (quoted after Sir Martin Rees). There is no clear answer to how we best try to detect life but investigating which planets can be expected to have favourable conditions for life seems to be a good first step. Here we review physical ideas about planetary interiors relevant for exoplanet habitability.

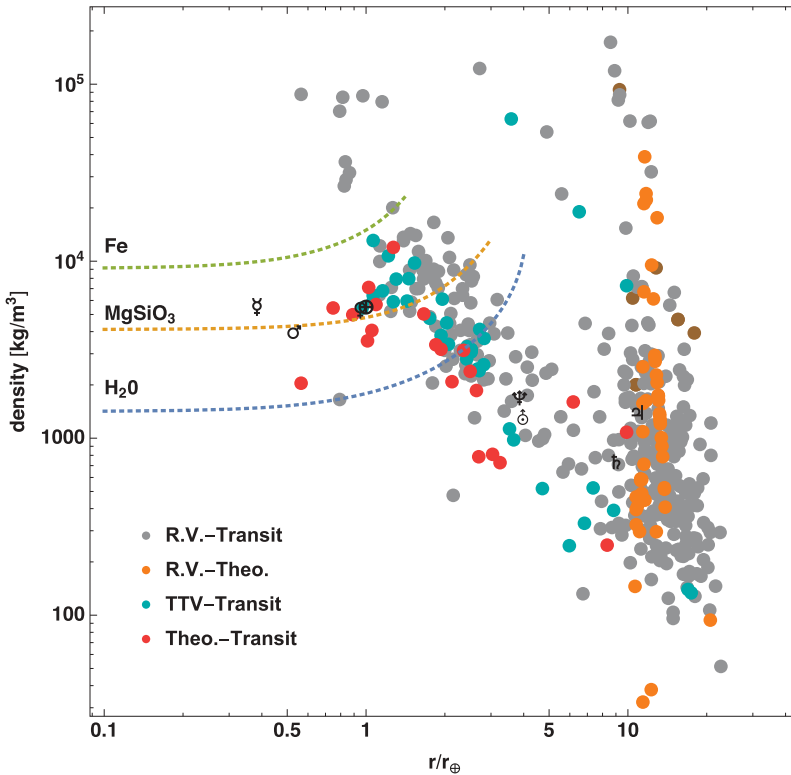
Data on single exoplanets is extremely limited compared to the information on the planets of the Solar System that has been obtained since the advent of space exploration of the Solar System in 1959. But in contrast to the wealth of data on the eight Solar System planets, the large number of detected exoplanets has the advantage of being well suited for comparative planetology. In the Solar System, half of the planets are of terrestrial type, with a core consisting mainly of iron and an envelope of rocks primarily composed of silicate minerals. The other four are giant planets that are built essentially of the most abundant volatile elements in the Solar System H, He, C, N and O. Despite the lack of detailed information on individual exoplanets, a much larger diversity has already been discovered among exoplanets than among Solar System planetary objects, awaiting more profound investigation in the next decades [5].

Most exoplanets have been detected by the *transit method* [6]. When a planet crosses the direction from the observer to the star during its orbital motion, it blocks part of the light emitted by the star. A sensitive photometer can measure this dip in the light curve of the star. As an example, Jupiter can block about 1% of the light of the Sun, the Earth only less than 0.01% as seen from outside the Solar System. If the star’s radius can be determined, the radius of the planet can then be inferred. Observations of small variations in the radial velocity of a star from Earth by spectroscopy also allow exoplanets to be detected. This *radial velocity (RV) method* led to the first detection of a planet around a main sequence star (51 Peg) in 1995 [7] and is the second-most common method for exoplanet detection. It is based on the principle that a star with an exoplanet will orbit around the common center of mass and therefore will move periodically to and from us, if the orbital plane is not perpendicular to the line of sight (‘face on’ orbit). This motion can be detected through blue and redshifts of the stellar spectrum. As an illustration of the radial velocities induced in stars due to the orbital motion of a planet, Kepler’s laws show that the Sun’s radial velocity is 12.4 m/s due to Jupiter

and 9 cm/s due to the Earth. The latter value is about the precision that can be reached by the newest generation exoplanet hunter spectrometer ESPRESSO, recently installed at the European Southern Observatory (ESO) [8]. The mass of the exoplanet can be determined from radial velocity observations if the orientation of the orbital plane with respect to the direction to the Earth is known, which is for example the case for transiting planets. In both methods, the exoplanet is detected indirectly through observations of variations in the light of the star. Before an observation can be shown to indicate convincingly the presence of a planet, other causes of variations have to be excluded, such as stellar variability due to intrinsic oscillations or due to darker regions on the surface caused by the stellar magnetic fields (similar to sunspots).

Both mass and radius are known of transiting exoplanets for which also radial velocity observations have been performed. From those two basic quantities the mean density can be estimated and constraints on the composition and interior structure can be inferred that allow a classification of the planet. Unfortunately, observational requirements lead to a limited overlap in sets of planets observed by either transits or radial velocity (see [Figure 1](#) for a representation of the mean density as a function of radius for identified exoplanets). For example, only 117 out of a total of 1184 exoplanets with radius less than 2 Earth radii have known mass and radius (data from [exoplanet.eu](http://exoplanet.eu), consulted on 27 May 2019), and 77 of them have an error on the mass larger than 10%. Radial velocity observations need sufficiently bright stars or large telescopes, whereas transiting planets are more easily observed for small, not very luminous stars since the planet will then block relatively more of the stellar radiation travelling towards Earth. Both methods also have a bias towards detecting planets close to the star since a decrease in orbital distance increases the relative size of the disk of the planet in front of the star and decreases the orbital period. Therefore, it also increases the radial velocity of the star and decreases the time required to observe the star-planet system for at least a few orbital cycles. Large and massive planets also are easier to detect as they increase the relative size of the planet and increase the velocity of the star around the common center of mass.

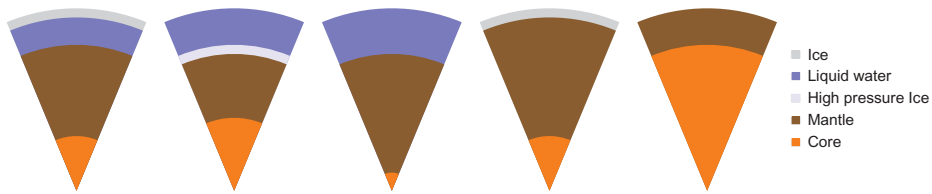
For those reasons, the first exoplanets detected were giant planets close to their host star. Evolutions in observational techniques, such as more precise RV measurements, and new developments with dedicated space missions have made the detection of smaller planets close to the host star possible, in particular planets that could have a rocky composition as the terrestrial planets of the Solar System. A remarkable example is the TRAPPIST-1 planetary system in which seven Earth-sized exoplanets transit the nearby ultracool dwarf star TRAPPIST-1 [9]. Future missions, in particular the ESA M3 mission PLATO foreseen to be launched in 2026, will enable to detect and



**Figure 1.** Mean density of exoplanets with radius and mass determination. Dashed lines indicate theoretical predictions for a purely iron,  $\text{MgSiO}_3$  and  $\text{H}_2\text{O}$  composition. Planets of most interest for habitability are situated between those lines. At the high radius end of the figure are the gas giant planets. Intermediate between the smaller planets considered here to be those with a radius  $\leq 2R_{\text{Earth}}$  is a group of planets with density decreasing with increasing radius and increasing amount of H-He envelope. Also indicated are the eight planets of the Solar System. Colours indicate the data source given for mass and radius. ‘R.V.’: radial velocity method, ‘Transit’: transit method, ‘TTV’: transit timing variations, ‘Theo.’: theoretical predictions.

characterize Earth-sized planets around solar-type stars at a similar distance as the Earth.

Besides planets similar to the Solar System planets, many, if not most, of the identified planets have significantly different characteristics (see [Figure 2](#) for a schematic representation of plausible planetary interiors). Low density ‘Hot Jupiters’, hotter and larger than was expected for giant planets consisting mainly of H and He, have been observed [e.g. 7]. Also extremely high-density planets exist, which seem to be made up almost exclusively of dense material like iron [10]. Maybe those planets formed like that, or lost their less dense outer layers, possibly due to collisions in the early stages of planetary formation or due to thermal effects. Quite striking is that about half of the identified exoplanets have a radius larger than the Earth (the largest Solar System terrestrial planet) but smaller than the ice giant Neptune, which is smaller than Uranus and has



**Figure 2.** Schematic representation of some plausible but non-exhaustive interior structures of exoplanets with radius up to 2 Earth radii. Atmospheres are not indicated.

a radius of about 3.9 times the radius of the Earth. Their observed mean density ranges from below that of Solar System gas giants to above that of Solar System terrestrial planets. These data demonstrate that different classes of exoplanets exist with different compositions and structures intermediate between terrestrial planets, which have a rocky envelope and an iron core, and ice giants composed primarily of hydrides as water, ammonia and methane. Before the detection of exoplanets, it was thought that there should be a gap in radius between terrestrial planets and planets that can accrete and retain a significant mass fraction of gas. Initial studies show that planets with a denser composition like terrestrial planets dominate at radii below about 1.5 to 2 times the Earth's radius and that larger planets are capable of capturing and maintaining a gaseous H-He envelope [11–14].

We here focus on exoplanets that might have the right properties for being habitable, meaning that they have physical conditions that are thought to be required to support life. We do not consider giant planets and restrict our attention to planets with a radius less than about 2 Earth radii that might have liquid water at the surface or below an ice crust, since water is essential for life as we know it. We review physical ideas about planetary formation, interior structure and evolution, and explore physical characteristics that can make a planet habitable. In [Section 2](#), we present an overview of current ideas about planetary formation and differentiation. [Section 3](#) introduces the general physical principles of planetary interior and evolution, and explains mass-radius relations and their use as an indication of the composition of the planet. The basic ideas are further developed and applied in [Sections 4](#) and [5](#), respectively, to the core and the mantle. Questions of habitability are addressed in [Section 6](#). Conclusions and perspectives are presented in [Section 7](#).

## 2. Formation and differentiation

### 2.1. Planetary formation processes

Planets form from a circumstellar disk of gas and solid dust particles shortly after the formation of the star. These disks, with typical sizes between 100 and 1000 AU [15], are commonly observed around young

stars. The dust grains contribute only  $\sim 1\%$  to the mass of the disk and typically have sizes of a few micrometer or less in the interstellar medium. The gas in the circumstellar disk quickly disperses in several million years due to photo-evaporation and viscous spreading. Remarkably, planetary formation processes are able to build planetary objects from a substantial part of the tiny dust grains in the protoplanetary disk in a few million years, an increase in mass and size by about 40 and 15 orders of magnitude. It is perhaps no surprise then that the formation process is not yet well understood. The discovery of many exoplanetary systems, often with orbital and physical characteristics quite different from the Solar System, has raised further questions and challenges, but at the same time has given more insight into the mechanisms and diversity of planetary formation. In current models of planetary formation, up to five different accretion stages can be distinguished: (1) coagulation of dust particles to sizes of several centimeters, (2) formation of kilometer-sized objects or planetesimals, (3) growth of planetesimals to objects with radius of about 1000 km (protoplanets or planetary embryo's) or more, (4) accretion of primordial gas for massive planets, and (5) possible growth by massive collisions of protoplanets due to gravitational instability of the planetary system after dispersal of the gas.

In the first stage, dust grains grow through collisions and chemical bonding in the dense disk. This process is size limited. Once the objects reach sizes of centimeters to 1 meter, they would fall onto the star due to drag from the gas in a time shorter than the collisional growth time. In addition, mutual collisions for particles of those sizes, especially in the inner part of the disk, become destructive through erosion or fragmentation. In the second stage, particles overcome this *metre-size barrier*, most likely through disk instabilities that increase the concentration of particles, although the details of these mechanisms remain uncertain [e.g. 16–21]. Consider for example that the particle density is locally higher. The particles there accelerate the gas more than elsewhere so that they feel a weaker headwind from the gas. They therefore drift to the star more slowly than average particles. Particles further away from the star in a neighbouring region with a lower particle density drift faster to the star and can join the denser region, increasing the particle density there. Those regions will eventually contract under self-gravitation and form planetesimals. This completes the second step. As an example, in our Solar System, many asteroids are planetesimals that have not evolved to a planetary size. Several mechanisms have been proposed for the third and last stage to form objects of planetary size. Most likely, individual planetesimals grow through accreting small particles (or pebbles) in the disk rather than by gravitational interactions with other planetesimals, in particular for the growth to large planets like super-Earths or cores of gas giants [22,23]. For the Solar System, this process can lead to tens of Mars-sized planets in the

inner Solar System and to planets with a mass of more than ten times the Earth's mass beyond the snow line, the distance from the star where water vapour changes to ice (temperature of 160–170 K [24]), in a few million years.

In a fourth stage, heavy planets can accrete H and He from the gas if they are formed before dispersal of the gas. When the gas forms the main part of the final planetary mass, which is thought to occur for planets with an initial mass of more than about fifteen Earth masses [24–26], the planets are called gas giants. If the planet consists of only a limited amount of H and He gas compared to the mass of rocks, iron and ices, it is commonly referred to as a Neptune-like planet. Statistical survey studies indicate that there is a bimodal distribution in radius of exoplanets smaller than gas giants, indicating two different populations: (1) planets with a large mass fraction of volatile elements (in particular hydride ices) typically have radii larger than about 2 times the radius of the Earth, and (2) rocky exoplanets have a radius usually smaller than  $\sim 1.5R_{\text{Earth}}$  [11–14]. Those rocky planets accreted much less volatiles, and may, as did the Solar System terrestrial planets, have only completed their formation after the dispersal of the gas from the disk. Somewhat confusingly, planets with a radius between 1 and 2 times the radius of the Earth, are also often called super-Earths [27]. We here restrict that term to those exoplanets that have a mainly rock-iron composition.

When the gas is eventually removed from the disk, the system of protoplanets can be densely packed and become unstable due to the absence of the damping effect of the gas on the orbital eccentricities [28]. Collisions between the protoplanets can lead to further growth and a limited final number of planets. Large impacts basins on Mercury, Mars and the Moon are witnesses in the Solar System of such large impacts, as is the formation of the Moon itself, thought to be due to a Mars-sized object impacting the early Earth about 50 million years after the formation of the Sun. Also the large obliquities (the tilt of the planets with respect to the orbital plane) of Neptune and Uranus are likely indicators of large impacts.

As mentioned in the introduction, we do not consider gas giant exoplanets since they don't have liquid water close to their surface and are not considered habitable. Besides terrestrial planets, which have not been able to capture or keep an H and He atmosphere from the protoplanetary disk, Neptune-like planets, which do have a primordial atmosphere can be of interest for planetary habitability. They form within the lifetime of the disk and can migrate inward to the star through interaction with the gas. Those exoplanets are often observed in orbits close to the star, where strong stellar irradiation can completely or partially remove their primordial atmosphere.



Beyond the snowline of a protoplanetary disk,  $\text{H}_2\text{O}$  exists in solid form and planets forming there can accrete large quantities of water. In the Solar System, all the moons of the outer Solar System planets, with the exception of Io, and most likely all Trans-Neptunian objects, consist of significant amounts of water. For example, the largest moons of the Solar System Ganymede and Titan, both several percent larger in radius than the smallest planet Mercury, have mean densities intermediate between water ice and rocks, suggesting that they consist of similar quantities of water-ices and rocks. Planetesimals in the inner disk well within the snow line are expected to be very dry, as is thought to be the case for those planetesimals forming the terrestrial planets of the Solar System [29]. Although planets forming from such planetesimals will be dry (drier than the Earth), a wide range of possible water abundances of several orders of magnitude, both below and above Earth's water content, is expected for terrestrial planets because of radial mixing with planetesimals from closer to or beyond the snowline [e.g. 29,30].

## 2.2. Planetary differentiation

When many particles are brought together from an extended region in the disk into the much smaller volume of a planet, the gravitational energy decreases by a value of the order of  $GM^2/R$ , where  $G$  is the universal gravitational constant, and  $R$  and  $M$  the radius and mass of the planet, respectively. If all gravitational energy is converted into heat, a planet would reach temperatures of  $GM/(RC_p) \approx 40000\text{K} \left(\frac{M}{M_\oplus}\right)^{2/3} \left(\frac{\bar{\rho}}{\bar{\rho}_\oplus}\right)^{1/3}$ , where  $C_p \approx 1500 \text{ J/K/kg}$  is used for the specific heat of the planet,  $\bar{\rho}$  is the mean density of the planet, and subscript  $\oplus$  indicates values for the earth. The temperature profile in the forming planet depends on the ratio between the energy delivered to the surface region and that delivered by large impacts to the deeper interior. If most energy is received by the surficial layers, the deep interior is expected to be colder than the outer layers [31]. Collisions between protoplanets and large impacts at the final stage of formation of rocky planets will bury enormous amounts of heat deep inside the planets. Impact simulation studies indicate that the interior temperature increases by several thousands of degrees for a planet like the Earth [32]. Radiative cooling of a hot planetary surface can strongly reduce the temperature of the surface layers within the formation time scale of the planet, but will not be able to cool the deep interior of a planet because the cooling of a planet's interior is a slow process (see below). A hot interior is therefore expected for planets shortly after formation. In the Solar System, all terrestrial planets currently have a hot interior of a few thousand K.

The temperature increase from accretion is expected to be sufficient to partially melt a planet with a rock-iron composition and to form a magma ocean. The decay of the short-lived radioactive isotopes  $^{26}\text{Al}$  with a half-life of 0.72 My and  $^{60}\text{Fe}$  with a half-life of 1.5 My can contribute to the melting for rapidly forming planets or smaller bodies such as Vesta [33]. Iron droplets and metallic siderophile compounds accumulate at the base of the magma ocean in large ponds and can sink to form an iron core by processes such as dyking, diapirism, and percolation [34]. Likewise, the heat of formation can lead to separation of rock from water and ice as has happened in some of the large icy satellites of the Solar System such the Galilean satellites Europa and Ganymede [35]. During differentiation, additional gravitational energy is released and converted into heat, further increasing the internal temperature and facilitating separation of iron, rock and water. But a higher mass and a resulting higher energy of accretion is not a guarantee for differentiation since planetary materials become miscible at high temperatures because the higher entropy of the mixed state leads to the lowest Gibbs energy for sufficiently high temperatures [36]. Recent ab initio calculations show that Fe and MgO are fully mixed at  $\sim 4000$  K at zero pressure and at  $\sim 7000$  K at the pressure of the boundary between the mantle and the core of the Earth. Differentiation requires lower temperatures for iron to be able to separate from the molten rock-iron mixture. An alternative pathway to form coreless planets exists if all the metallic iron in the accreting material is oxidized for example by water [37]. Iron will then form iron oxide and bind with the silicate minerals of the planetary mantle.

### 3. Basic equations of interior structure and evolution

The bulk of the hot planetary interior cannot support shear stresses on very long timescales and behaves as a viscous fluid. The characteristic time scale that separates elastic from viscous liquid behaviour is most easily given by the Maxwell time defined as the ratio of viscosity to shear modulus (rigidity). For the Earth's mantle, the Maxwell time is of the order of thousand years. Because of the fluid-like behaviour on geological timescales planets are close to hydrostatic equilibrium. For the Earth for example, the deviation in polar flattening from that of a hydrostatic mass distribution as a result of rotation is less than 1%. Hydrostatic equilibrium expresses that the downward gravitational force is balanced by the upward differential pressure force at any point in the planet at a radial distance  $r$  to the planet's mass center:

$$\frac{dP(r)}{dr} = -\rho(r)g(r), \quad (1)$$

where  $P(r)$  is pressure,  $\rho(r)$  mass density, and  $g(r)$  gravity. We assume here that the planet is spherically symmetric. Rotational and tidal deformations, which may lead to significant deviations from sphericity, can be calculated afterwards for any spherical reference model. By assuming the density to be homogeneous, a lower limit to the pressure in the center of planets can be calculated from Equation (1) as  $P_c = 3GM^2/(8\pi R^4)$ , where  $M$  and  $R$  are the mass and radius of the planet, respectively. For the largest super-Earths with a radius of 2 Earth radii and a chosen mass of 10 Earth masses [e.g. 27], the central pressure can then be estimated to be about 2 TPa. This equation for a homogeneous planet underestimates the central pressure by a factor two for the Earth due to material compressibility and even more for more massive planets (see Figure 3), so that pressures up to at least 10 TPa have to be considered, well beyond the pressure in the centre of the Earth of 364 GPa.

According to Newton's theory of gravitation, gravity and density satisfy Poisson's equation everywhere in the planet:

$$\frac{dg(r)}{dr} + \frac{2}{r}g(r) = 4\pi G\rho(r). \quad (2)$$

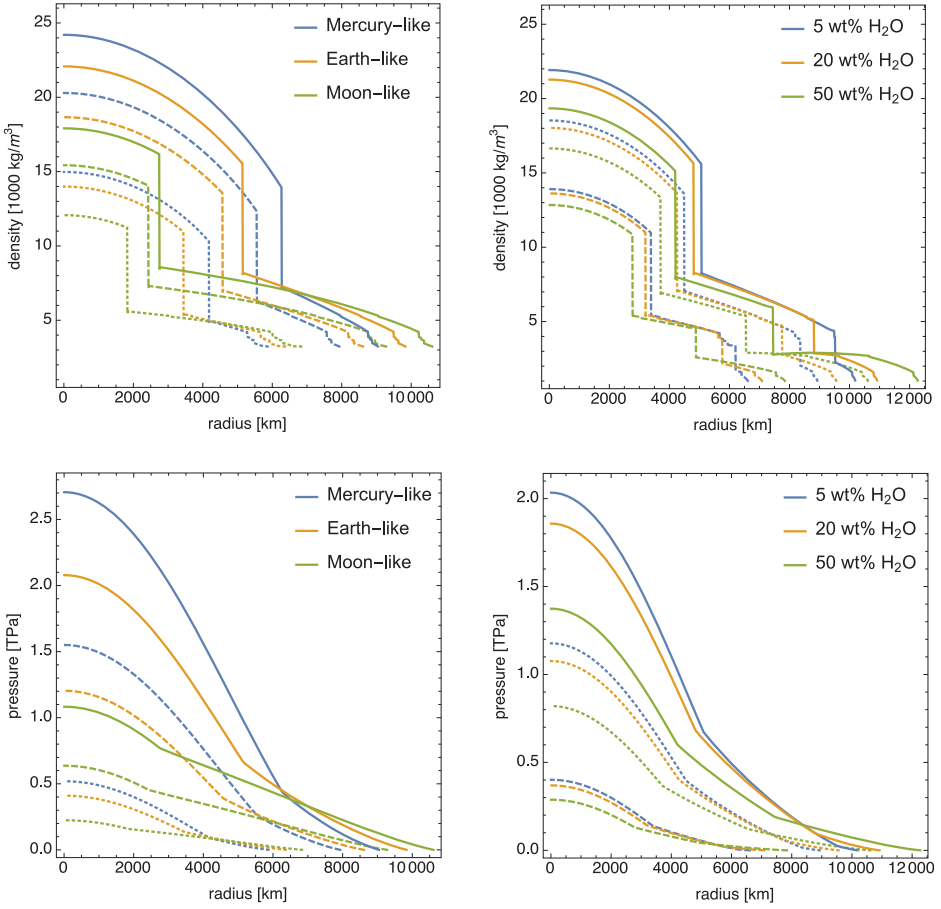
Besides pressure and gravity, Equations (1) and (2) depend on the density. A third equation is therefore needed to solve for these three variables. This is given by an equation of state (EoS) specifying how density depends on pressure, temperature, and composition. The temperature in the planet depends on the heat sources and on the way heat is transported.

Energy is transported in planets that dominantly consist of iron, rocks and ice, in solid or liquid form, through conduction and convection. For a layer in which heat is transported by conduction, the temperature is related to the heat flux by

$$q(r) = -k(r)\frac{dT(r)}{dr}, \quad (3)$$

where  $k(r)$  is the thermal conductivity. Conduction is not very efficient in cooling the deep interior of a large planetary body. On a timescale of 10 Gyr, rocky bodies with a radius larger than 1000 km cannot efficiently cool down to their centre by conduction. Effective cooling is only possible if heat is transported by convection in a significant part of the planet.

Convection transports heat through advection and is driven by buoyancy, the net gravitational force downward on a denser material than the environment and upward on a less dense material. The differences in density in a planet are due to either thermal or compositional effects. In plane layers, these situations give rise to Rayleigh-Bénard and Rayleigh-Taylor instabilities, respectively. The occurrence of convection depends on the value of the Rayleigh number  $Ra$ , which is defined for purely thermal convection by



**Figure 3.** Density (top) and pressure (bottom) profiles of super-Earth exoplanets and water-rich planets with masses equal to 1 (dashed), 3 (dotted) and 5 (continuous) Earth masses. The super-Earths have a bulk iron wt% of 35 (Earth-like), 5 (Moon-like) and 63 (Mercury-like), a silicate magnesium number of 0.9 and no H<sub>2</sub>O. For the water-rich planets, bulk H<sub>2</sub>O wt% of 50, 20 and 5 are considered. The 50 wt% H<sub>2</sub>O planets have 17.5 wt% iron and 32.5 wt% rocks (Mg number of 0.9), those with 20 wt% H<sub>2</sub>O have 28 wt% iron and 52 wt% rocks (Mg number of 0.9), and planets with 5 wt% H<sub>2</sub>O planets have 32.25 wt% iron and 61.75 wt% rocks (Mg number of 0.9).

$$\text{Ra} = \frac{\rho g \alpha \Delta T d^3}{\kappa \eta}, \quad (4)$$

where  $\alpha$  is the thermal expansivity,  $\Delta T$  the non-adiabatic temperature difference between the top and the bottom of the layer,  $d$  the thickness of the layer,  $\eta$  the viscosity and  $\kappa = k/(\rho C_p)$  the thermal diffusivity, with  $C_p$  the specific heat at constant pressure. Convection occurs when Ra is larger than a critical value and Equation (4) shows that this happens more easily for larger buoyancy ( $\rho g \alpha \Delta T$ ) and a thicker layer. Viscosity and thermal diffusivity hinder convection by restricting motion and

decreasing the density contrast with the environment. The Rayleigh number can be interpreted as a ratio between the thermal diffusion cooling timescale ( $\propto d^2/\kappa$ ) and a timescale of motion at the Stokes velocity ( $\propto \rho g \alpha \Delta T d^2/\eta$ ). If the cooling timescale is long with respect to the timescale of motion, convection can develop, otherwise thermal equilibration will be too fast and/or the Stokes velocity too slow for convective motions to develop [38].

Convection is an efficient heat transport mechanism and therefore leads to a temperature gradient that is very close to the adiabatic gradient. The adiabatic gradient is defined as  $(dT/dr)_{ad} = -\rho g T/C_P$  and is the minimally required temperature gradient for convective instability to occur in inviscid fluids (Schwarzschild criterium). In a planet, the temperature difference between the hot bottom and the colder top of a layer can be much larger than what is expected for an adiabatic profile (for the Earth a factor of about three for the silicate layer). Thermal boundaries will therefore form at the top and bottom in which the temperature drop is much steeper and the convective velocities are zero. The heat flux is transported by conduction through the boundary layers and the boundary layers both drive the convection and determine the heat flux out of the mantle.

Heat can be produced in exoplanets by various processes. Besides the heat generated by the accretion process, decay of radio-active isotopes and heating associated with tidal motions are the main heat sources in the Solar System planets and satellites and are undoubtedly also important for exoplanets. Tidal heating strongly increases with decreasing distance  $a$  to the central object as  $a^{-6}$  and increases linearly with the mass of that tide-raising body. In the Solar System, it is important for satellites close to a large planet. As an example, on Io, the Galilean moon of a similar size and distance to Jupiter as our moon to the Earth, tidal dissipation leads to massive volcanism [39]. Likewise, for exoplanets on close-in eccentric orbits, tidal dissipation can be expected to be a more important heat source than for the Earth, and might even be the dominant energy source as in Io [40]. Other heat sources for exoplanets include stellar irradiation, as invoked for close-in hot Jupiters [e.g. 41], and heating by magnetic induction in close-in super-Earths and giant exoplanets [e.g. 42,43].

The abundance of radio-active isotopes is determined by their abundance in the proto-planetary disk and by the accretion processes. Both factors are difficult if not impossible to constrain. Even for the terrestrial planets of the Solar System other than the Earth, the abundance of radio-active elements is not well known. The main isotopes able to provide energy over the entire life of a planet are  $^{235}\text{U}$ ,  $^{238}\text{U}$ ,  $^{232}\text{Th}$ , and  $^{40}\text{K}$ . The first three are refractory elements, meaning that they belong to the elements that remain solid up to high temperatures in the proto-planetary

disk, whereas  $^{40}\text{K}$  is more volatile and requires lower temperatures to occur in solid form. Potassium is therefore expected to be depleted in planets that form in the inner zones of the proto-planetary disk compared to those that form in the outer disk. Thorium has a half-life about equal to the age of the universe, whereas the other three have a half-life of the order of 1 Gyr. In the first phases of evolution, the radiogenic heat produced is largest and can be sufficient to heat up planets by several 100 K, depending on the concentration of radio-active isotopes and the efficiency of heat transport (see [Figure 7](#)). For the Earth, the radio-active energy production is now a factor 4 smaller than at the time of formation and similar factors likely apply to exoplanets. Radio-active heating is more important for larger, more massive super-Earths that have a large volume-to-surface and can delay the final slow cooling phase by several Gyr [e.g. 44,45].

The total heat flux  $L$  out of a planet can be well approximated by the sum of radioactive energy production  $\epsilon_R$ , the energy production by tidal dissipation  $\epsilon_T$  and the loss of thermal energy:

$$L = \int_0^M \left( \epsilon_R + \epsilon_T - C_V \frac{dT}{dt} \right) dm$$

This heat flux is much smaller than the reflected and reradiated stellar flux for the terrestrial planets of the Solar System, implying that their surface temperature is mainly determined by the received solar radiation and not by their interior. For example, for the Earth, the received flux of the Sun is about 2750 times larger than the internally produced heat flux. A low internal to stellar received heat flux may also apply to exoplanets with a radius below 2 Earth radii after an initial fast cooling phase (note that the known semi-major axis for all but one of those planets known to this date is below 1 AU), although tidal heating might become dominant for very close-in planets.

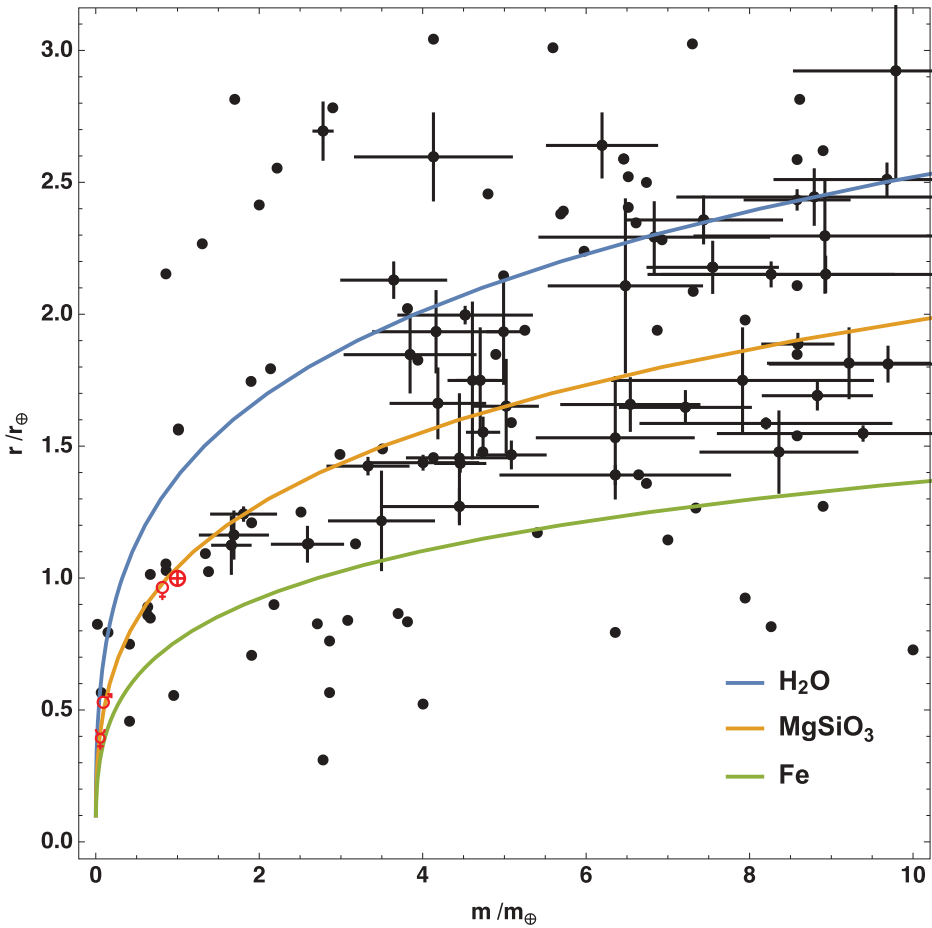
If radiogenic heating were the only heat source of the planet and were to supply energy at a constant rate, a balance between the internal energy source and the loss of energy through the surface could exist and a planet might keep a constant temperature. However, the radioactive isotopes decay so that the amount of heat-producing elements decreases with time. Since the internal heat produced in the planet declines with time, the surface heat flux also declines with time. The transported heat must then diminish, and therefore also the internal temperature must decrease with time. Planets therefore necessarily eventually cool. The terrestrial planets of the Solar System cool by some tens to about 100 K per Gyr and similar cooling rates are expected for terrestrial-type exoplanets after having reached the slow cooling regime (see [Figure 7](#)). In that regime, the mantle temperature and viscosity take values that facilitate the removal of the heat produced plus part of the

primordial heat. If less heat would be transported than is created internally, the mantle temperature would rise, viscosity would decrease, and more heat would be transported until at least as much heat would be transported as is produced.

Investigations of the interior structure and evolution of potentially habitable exoplanet use the same basic physical principles as for the terrestrial planets and satellites of the Solar System, but face additional problems. Not only are the observational data about individual exoplanets extremely limited compared to data about planets and moons of the Solar System, but the physical conditions in exoplanets might also be very different from those for which physical data on planetary materials are available. For example, pressure (Figure 3) and temperature can be significantly higher than in Solar System planets and the behaviour of relevant materials at these conditions is often not known from experiments or theory. In addition, thermodynamic and transport properties are not well constrained for compositions differing from those in the Solar System and quite likely even unknown stable compounds can occur in exoplanets. These drawbacks can also be turned into an advantage since exoplanets can be considered a laboratory for investigating the behaviour of material at extremely high pressure and temperature.

Many studies have addressed the possible interior structures of super-Earth exoplanets by solving the above structural equations for chosen planetary compositions [27,46–50]. These studies assume that the exoplanet is differentiated into a core consisting essentially of iron and a rocky mantle, and possibly a H<sub>2</sub>O layer. Figure 3 shows density and pressure profiles for 3 different planet masses and different compositions for both dry super-Earth planets and planets with a significant H<sub>2</sub>O layer. The different layers are easily identified in the density profiles, as are the increases in density due to compressibility and phase transitions. The density can reach up to more than 24,000 kg for a 5 Earth mass super-Earth with a large core, about 70% higher than in the centre of the Earth. The pressure can increase to about 2.7 TPa for those planets, more than 7 times higher than in the centre of the Earth.

A basic and much used result of interior structure models are theoretical mass-radius relations, which can be compared with observational mass and radius data to estimate the interior composition and structure of an observed planet (Figure 4). Since planetary materials are compressible, the radius  $R$  of a planet of fixed composition increases more slowly than  $M^{1/3}$  (see Figure 4), as would be the case for incompressible material, and the mean density of the planet increases (Figure 1). With increasing mass and increasing internal pressures, the material compressibility decreases and therefore the radius increases more slowly with



**Figure 4.** Mass-radius relation for exoplanets with mass smaller than or equal to 10 Earth masses. Three end-member compositions are considered: pure iron,  $\text{MgSiO}_3$  and  $\text{H}_2\text{O}$ . Error bars are indicated for planets with errors  $< 25\%$ . Temperature effects have been neglected [e.g. 47,50].

increasing mass (Figure 4) and the mean density increases faster (Figure 1).

Although the comparison between theoretical mass-radius relations for specific compositions and interior structures and the observed mass and data can be used to distinguish between a primarily rocky or gaseous planetary composition, there is a large degeneracy mainly because many different compositions and structures can lead to the same mean density [e.g. 50,51]. On the most basic level, one might distinguish between planetary material as iron/metals, rocks, ices and H/He gas (in order of decreasing density), but their relative proportions cannot be determined based on the mean density only, in particular for planets of intermediate mean densities, which might consist of a range of materials from the least to the most dense. As an example,



a planet with the mass and radius of the Earth, might have a substantial water/ice layer and a larger iron core than the Earth [52]. If an exoplanet has an extended atmosphere of mainly H and He, the estimated composition of the planet beneath the atmosphere would be drastically different from the composition of the planet assumed to have no atmosphere. The often large errors on mass and radius (see Figure 4) further add to the degeneracy. The degeneracy for potentially habitable planets can be well illustrated by the TRAPPIST-1 planets. For example, planet d of the TRAPPIST-1 planetary system with an uncertainty on the mass of 45% (estimated mass of  $0.33 \pm 0.15M_{\text{Earth}}$ ), an estimated radius of  $R = 0.772 \pm 0.030R_{\text{Earth}}$  and an expected surface temperature in the right range for liquid water to occur, could consist of an iron core relatively about as large as that of the Earth and a rock mantle, but its mass and radius can also be satisfied by a planet made of rocks and  $\text{H}_2\text{O}$ , with the  $\text{H}_2\text{O}$  occupying up to 80% of the volume [53]. With an improved mass accuracy by a factor 4 and a 20% smaller uncertainty on the radius (mass of  $0.297 \pm 0.037M_{\text{Earth}}$  and radius  $R = 0.784 \pm 0.023R_{\text{Earth}}$  [54] more stringent constraints can be derived and the planet must have a significant water volume fraction of at least about 20% [55]. With perfectly known mass and radius, the ranges of volume fractions could be determined for iron (0–20%), rock (0–60%) and  $\text{H}_2\text{O}$  (45%–75%) [55], assuming that the planet does not have a significant atmosphere. If an exoplanet can be assumed to consist of an iron core and a silicate mantle only, accuracies of 2% and 10% on radius and mass expected by the PLATO mission, allow estimating the core mass fraction with an uncertainty of about 30% [56].

#### 4. Core

The cores of the terrestrial planets in the Solar System are mainly made of iron. Not only is iron the only heavy element abundant enough in the Solar System, it also has about the right density and elastic properties that correspond with seismic observations of the Earth [57]. For the other Solar System terrestrial planets, where seismic data are not yet available, only an iron rich core can explain the internal mass distribution deduced from their moment of inertia. As explained above, it is thought that, much like terrestrial planets and some moons of the Solar System, exoplanets also have an iron-rich core below a silicate or silicon-carbide shell if pressure and temperature conditions allow for their immiscibility.

The size of the core depends mainly on the composition of the planetary nebula, the formation processes, and also on the redox conditions of the environment in which planets formed [e.g. 58]. As an example, among the Solar System planets, Mercury has a core radius of about 80% relative to the planet radius whereas the relative core radius of the other terrestrial planets like Mars is about 50% [59–61]. Mercury probably lost part of its mantle due

to giant collisions [62,63] and formed under reducing conditions [64] with a relative large core and with an iron-poor mantle. Mars formed under more oxidizing conditions with a relative smaller core and with a mantle that has relatively about twice the iron-content of the Earth-mantle.

The pressure and temperature conditions and to a lesser extent the presence of other chemical compounds determine the phase and state of iron in the core [65]. The crystal structure of solid iron at the pressures of the smaller terrestrial planets in the Solar System is well established by means of experimental measurements to be in the fcc state inside Mercury, the Moon, and Mars and in cores of large icy satellites. For the core of the Earth, experimental measurements at the relevant pressures and temperatures are still challenging and cannot conclusively decide between the hcp phase and the bcc phase [e.g. 65]. Whether a phase is stable with respect to another is determined from its Gibbs free energy. At ambient conditions Gibbs energies are available for a large number of compounds and elements and can be calculated from equations of state and melting data at moderate pressures. In principle, Gibbs energies can also be computed from ab initio methods but those calculations are still quite challenging. At 0K, definitive answers about the stable phase of iron as a function of pressure can be obtained from ab initio enthalpy calculations [e.g. 66]. They show that hcp iron is the most stable phase at pressures in the Earth's inner core. At the high core temperatures, however, the most recent results favor fcc, bcc, or other close-packed structures over hcp iron [67–69]. For more massive planets, where the local pressure and probably also temperature in the core may significantly exceed those inside the Earth's core (Figure 3), experimental results are not available. Ab initio calculations at 0 K show that hcp iron remains the most stable phase up to 8 TPa, that the fcc phase is most stable up to 24 TPa, and that at 35 TPa there is a take-over by the bcc-phase [e.g. 66,70]. The phase of solid iron affects its thermoelastic and transport properties and to some extent the partitioning of light elements between its solid and liquid phase. Those properties influence the density of the core and the energy available to drive a core generated dynamo. Fortunately, the effects of the precise phase of iron in the core of a large exoplanet decrease with pressure [66] and are likely too small to be detectable in the near future.

Of more importance is the state of the core. It depends on the temperature in the core, which is extremely difficult to determine even for the Earth (uncertainty of about 1000 K), on the melting temperature of iron, which is known experimentally for pressures up to only  $\sim 200$  GPa [71], and on the light elements in the core. In the Solar System, the most likely light elements in cores of terrestrial planets are sulfur, silicon, oxygen, carbon, hydrogen, and magnesium [72,73], but their contributions are not well known, not even for the Earth. The presence of a light element decreases the melting temperature of the

alloy, and does so to first order stronger the more it partitions only into the liquid [74]. For example, oxygen that does not partition in solid iron at Earth core conditions decreases the melting temperature of iron significantly, whereas silicon that partitions almost equally in both the solid and liquid phase has almost no effect on the melting temperature of iron [e.g. 75]. At higher pressure than in the Earth's core, the effect of light elements on the melting curve is currently undetermined. Since all the factors determining the physical state of the core are unlikely to be known accurately for exoplanets, it may seem that the state of the core of super-Earths cannot be known. Nevertheless, *ab initio* calculations by [76] up to pressures of 1.5 TPa show that the melting curve of iron is steeper than a core isentrope at super-Earth pressures, suggesting that the core of massive super-Earths is solid if the temperature at the core-mantle boundary is below the melting temperature.

The state of the core of a planet, solid, liquid, or a liquid layer on top of a solid inner core as for the Earth has potentially observable consequences for exoplanets. If a planet has a liquid layer in its core, tides can be significantly larger than for a fully solid planet and consequently tidally driven dissipation is also larger and could have a significant effect on the planets internal and orbital evolution. Likewise, a planet with a liquid layer in its core could generate a magnetic field by dynamo action if the mantle allows the core to cool fast enough [e.g. 77]. Electrical resistance of the core alloy dissipates energy that must be supplied by the planet's internal heat, gravitational energy, radioactive heating and chemical energy. The efficiency of the core dynamo can be estimated from the entropy change in the core, which shows that significant amounts of entropy can be produced by exsolving light elements [73,78,79] and by the release of light elements when inner core formation occurs (the amount is larger if less light elements partition in the solid phase of the core) [e.g. 77]. Not all the energy and entropy produced inside the core is available for driving the dynamo. A significant amount of the produced entropy is lost by thermal conduction along the temperature profile of the core. The entropy of thermal conduction depends on the temperature gradient in the core and on the thermal conductivity of the core material. The larger the thermal conductivity the larger the amount of entropy lost by thermal conduction. The thermal conductivity inside the Earth's core is still highly debated and not known at higher pressures inside exoplanets. Both direct experimental measurements at Earth's core conditions [80,81] and theoretical calculations [82–84] are challenging and controversial. Further advances are needed since the value of the thermal conductivity at core conditions has important consequences on the thermal structure of the core, the evolution of an inner core, and the history of a global scale planetary magnetic field [e.g. 85]. The magnetic field is also relevant for planetary habitability since it may protect a possible atmosphere and life from stellar and cosmic winds.

## 5. Mantle

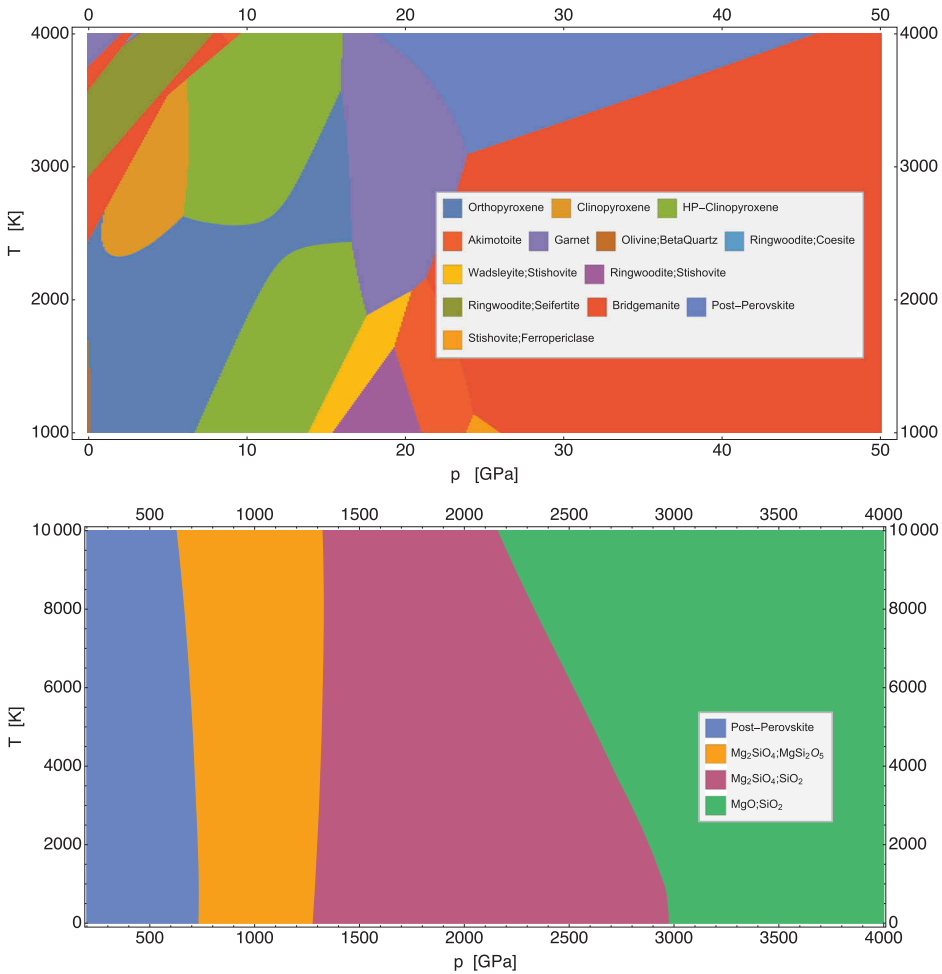
### 5.1. Composition and structure

The mantle of rocky planets is composed of various minerals. In water-rich exoplanets, the rock layer can be covered by a potentially thick H<sub>2</sub>O layer, as is the case in most large icy satellites of the Solar System. Most of the rocks in the terrestrial planets of the Solar System are silicate rocks, consisting of minerals that contain the silicate tetrahedron SiO<sub>4</sub> as their main building block, but a wide range of other minerals and iron oxides also occur. The major part of Earth's mantle minerals is made of Si, Mg, O, Fe, Ca and Al, together constituting more than 98% of the mass of the mantle [86]. Since those elements are the most abundantly produced non-volatile elements by stellar nucleosynthesis, they are thought to be the principal constituents of rocky mantles of most rocky exoplanets and the other terrestrial planets in the Solar System. The relative fractions of the different elements in the mantle depend in a complicated way on the composition of the protoplanetary nebula and on the formation processes of the planet, which together determine the bulk chemical composition of the planet, but also on the core formation process, which partitions elements between the mantle and the core. For the bulk Earth, the relative abundance of the chemical elements in refractory components – elements or minerals that have a high equilibrium condensation temperature above about 1300 K and condense first from a cooling protoplanetary nebula – is nearly equal to that of the Sun, suggesting that the refractory composition of a central star might be used as a first estimate of the bulk chemical composition of a rocky exoplanet [87–91]. However, even among the terrestrial planets of the Solar System, large differences in bulk compositions exist. This is illustrated most strikingly by the bulk iron fraction in Mercury, which is about twice that in the other terrestrial planets. Significant deviations also exist in the chemical composition of the mantles of the terrestrial planets of the Solar System. Mercury's mantle is almost devoid of iron and significantly more abundant in sulfur [64] than the mantles of the Earth and Mars, and the relative mass fraction of iron in the mantle of Mars is almost twice that of the Earth. A larger chemical diversity is expected for the rocky mantles of exoplanets, based on observed variations in the elemental composition of protoplanetary nebulae and central stars with respect to the Solar System, and on the complexity of planetary formation and differentiation processes. As an example of the variation in the mantle composition of exoplanets, mantles consisting primarily of silicon carbide and carbon rather than silicate materials are predicted for terrestrial-type exoplanets forming within the snow line in protoplanetary nebulae with carbon to oxygen ratios in excess of [92,93].

The stability and concentration of the minerals depend on the chemical composition and the local temperature and pressure. In particular, with

increasing pressure, the rather open structure of silicate minerals (essentially lattices of large O ions with interstitial other ions) becomes more closely packed and solid-solid phase transitions occur leading to denser minerals. On Earth, the density changes from the top to the bottom of the mantle by almost a factor two. The main upper mantle mineral olivine (chemical formula  $(\text{Mg}_x\text{Fe}_{1-x})_2\text{SiO}_4$ ) changes to wadsleyite at about 410 km depth and next to ringwoodite at a depth of approximately 520 km. The main phase transition in the mantle occurs at a depth of about 660 km, where ringwoodite becomes unstable and changes into bridgmanite  $(\text{Mg}_x\text{Fe}_{1-x})\text{SiO}_3$  and magnesiowustite  $(\text{Mg}_x\text{Fe}_{1-x})\text{O}$  and determines the boundary between the upper and lower mantle in the Earth at a depth of about 670 km. At a depth of about 2700 km, bridgmanite, which has a perovskite structure, changes into post-perovskite (ppv) [94], which is likely stable at least up to about 0.4 TPa [95]. The extent of the stability field of ppv depends on pressure, temperature, and composition. For pressures significantly larger than found in the Earth's mantle but possible in super-Earth planets, it has been shown that pure Mg-ppv undergoes series of successive transformations [95]:  $\text{MgSiO}_3 \rightarrow \text{Mg}_2\text{SiO}_4 + \text{MgSi}_2\text{O}_5 \rightarrow \text{Mg}_2\text{SiO}_4 + \text{SiO}_2$ . At pressures above about 2 TPa,  $\text{Mg}_2\text{SiO}_4$  dissociates into  $\text{MgO} + \text{SiO}_2$ . Figure 4 shows the  $(P, T)$  phase diagram for pressures between 0 and 4 TPa and temperatures up to 10 000 K for  $\text{MgSiO}_3$ .

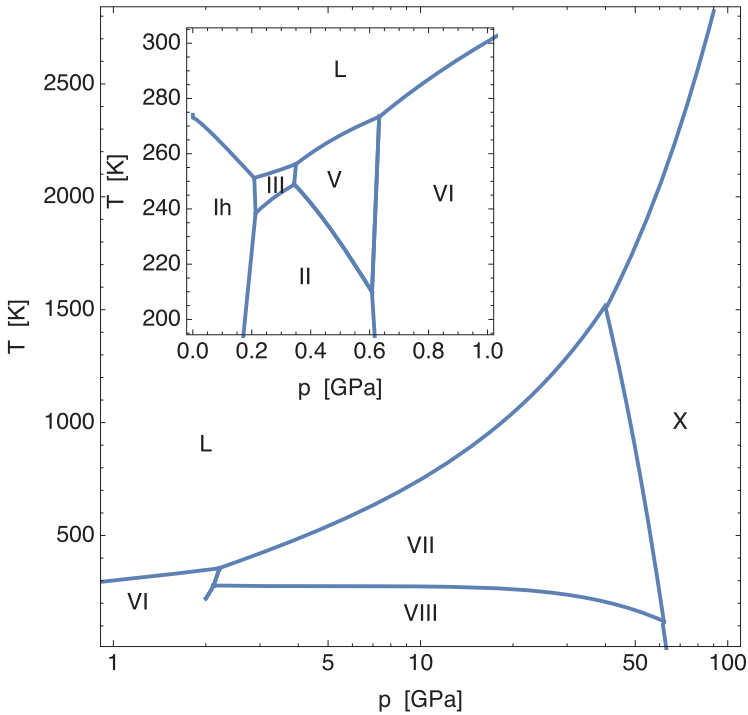
Compared to the silicate mantle, there are more phase transitions at relatively low pressures in an ice mantle because ice is more compressible than rock. Even in small Solar System moons and small exoplanets and exoplanetary moons, several of the phase transitions represented in Figure 6 occur. The most important property of the  $\text{H}_2\text{O}$  phase diagram for exoplanets is that the phase transition from water to an ice phase changes from endothermic at low pressures (below  $\sim 0.2$  GPa) to exothermic at higher temperatures. Therefore, the minimum temperature at which  $\text{H}_2\text{O}$  can be liquid ( $\sim 250$  K) is below the surface at a pressure of  $\sim 0.2$  GPa, which for a planet with the same surface gravity as the Earth is at a depth of about 20 km. As a consequence, even for a low temperature ice surface well below freezing, such as is the case for icy satellites of the Solar System, a liquid layer below an ice shell is possible if the temperature in the planet or moon increases sufficiently fast from the surface inwards in the low-pressure ice phase Ih to intersect the melting curve at pressures below  $\sim 0.2$  GPa. With increasing depth and pressure, the temperature profile will intersect again the melting, but this time at a transition to a high-pressure ice phase, and a planetary structure with consecutive layers from the surface of ice, water and ice occurs (see Figure 2). This situation is thought to exist in e.g. the largest moons of the Solar System Ganymede and Titan and in water-rich exoplanets.



**Figure 5.** Phase diagram for MgSiO<sub>3</sub> at planetary pressure and temperature conditions based on [96] for low pressures and on Umemoto:2017aa for higher pressures. The stability domains of the different minerals are indicated in colour.

## 5.2. Evolution

An initially hot and at least partially molten planet cools efficiently on a time scale of tens to hundreds of Myr until the whole silicate mantle (in particular the magma ocean) is solidified. Afterwards, the thermochemical evolution of rocky planets continues in a slower manner lasting billions of years, in which the planet can first heat up by several 100 K and eventually slowly cools (see example in Figure 7). During this second step of planetary evolution, heat, that is produced by radioactive decay, tidal dissipation and other internal heat sources, is transported from the interior to the surface by both convection and conduction. The heat transport by convection and the organization of convection in various convecting cells

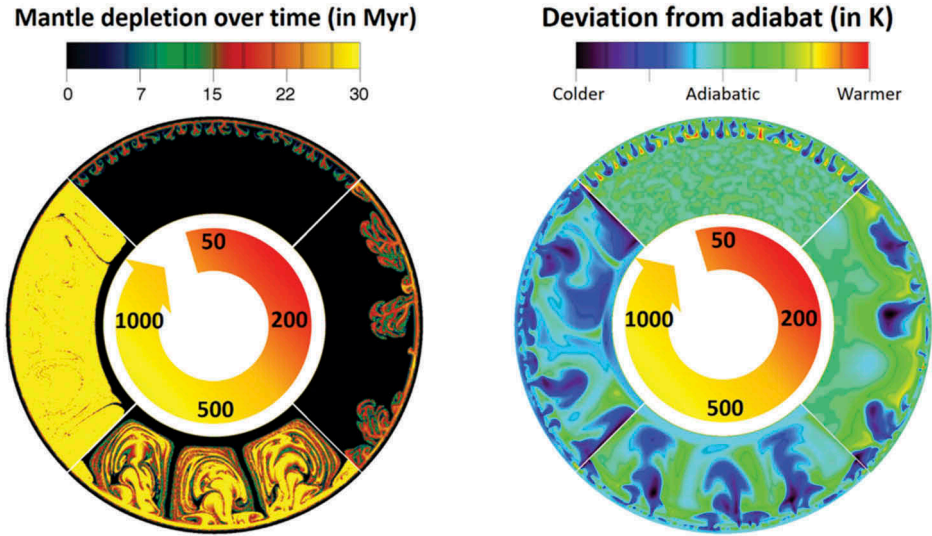


**Figure 6.** Phase diagram for  $\text{H}_2\text{O}$  at planetary pressure and temperature conditions based on [97]. The inset shows a zoom-in for pressures below 1 GPa.

depend strongly on the thermo-elastic and transport properties of the mantle material and on the occurrence of mineral phase transitions.

In particular viscosity strongly influences heat transfer in the mantle since it is a determining factor in mantle convection (see Equation 5). A high viscosity, as might be the case in a deep post-perovskite layer, will inhibit convection and reduce cooling of the deeper planet, which eventually may lead to heating of such a [44,98]. Evidently, also the thermal conductivity  $k$  influences heat flow. The thermal conductivity can increase by one to two orders of magnitude from the surface (e.g. for regolith or porous crust) to the core-mantle boundary of super-Earths [e.g. 99,100]. A low conductivity value at the surface leads to a strong isolation of the lithosphere and mantle from the surface, resulting in a heating effect of the upper mantle (and possibly melting). Unfortunately, for massive exoplanets, transport properties at relevant pressure and temperature conditions are not well constrained. Since their knowledge is often extrapolated from Earth-mantle relevant constituents, large uncertainties continue to exist in thermal evolution studies of exoplanets.

Also the exact composition, currently unknown for exoplanets, can have a significant effect on thermo-elastic and transport properties. Water in the mantle for example leads to a smaller rock viscosity (and hence faster convection



**Figure 7.** Mantle depletion and temperature evolution of the silicate mantle of a rocky exoplanet with a mass equal to that of the Earth and an iron bulk concentration of 35 wt% at four different times after formation: 50 Myr, 200 Myr, 500 Myr and 1 Gyr. The mantle becomes depleted due to partial melting and convective mixing. Its value is indicated as percent of extracted melt (between 0 and 30).

[101],), decreased melting temperature [102], and smaller melt densities [103,104]. Iron like water reduces the melting temperature [105] and mantle viscosity [106,107] but increases the melt density [103,104]. Carbon also strongly influences the interior structure and evolution of the mantle [108]. Silicon carbide for example shows decreasing melting temperatures for increasing pressures (measured up to 8 GPa [109],), and could influence rock melting at higher pressures. Carbon and SiC both have high thermal conductivity values, which could substantially change the transport of heat in the mantle [110].

During a first-order phase transition in the mantle, latent heat is either absorbed from the local environment (endothermic transition), or released to the mantle (exothermic transition) during the transformation of the mineral to the higher pressure phase. Since the density increases at a phase transition with increasing pressure, an exothermic (endothermic) phase transition corresponds to a positive (negative) Clapeyron slope  $dT/dP$ :

$$\frac{dT}{dP} = \frac{\Delta V}{\Delta S} = \frac{T\Delta V}{L}, \quad (5)$$

where  $P$  and  $T$  are the pressure and temperature of the transition,  $\Delta V$  and  $\Delta S$  are the volume and entropy change on transition, and  $L$  is the absorbed latent heat. Examples of exothermic phase transitions are the transformation of olivine into wadsleyite, of wadsleyite into ringwoodite, and of bridgmanite to post-perovskite (see Figure 5). The phase transition of



ringwoodite into bridgmanite and magnesiowüstite is endothermic (Figure 5). Cold descending slabs are accelerated at exothermic transitions and decelerated at endothermic transitions since the transition occurs at a lower (higher) pressure as a result of the positive (negative) Clapeyron slope. The displacement of the phase-transition boundary with respect to the surrounding mantle creates a local positive (negative) density anomaly for exothermic (endothermic) reactions promoting (opposing) the descending motion of the slab [e.g. 111]. Similarly, hot upwellings are decelerated at endothermic phase transitions since the higher temperature in the plume leads to a lower transition pressure, and hence denser material than in the colder surrounding material. This effect may have resulted in a layered mantle convection when the mantle was hotter during the early evolution of Earth [112]. Similar effects can occur in exoplanets (see Figure 5), also at higher pressure phase transitions, and influence the heat transport and mixing in the mantle, but require further investigation.

A specific characteristic of mantle convection on Earth is the occurrence of plate tectonics, a feature unique in the Solar System. We do not yet understand why plate tectonics did evolve on Earth and cannot predict its occurrence on exoplanets, but a basic requirement for plate tectonics is that the convective stress exceeds the yield strength of the brittle lithosphere and crust. Plate boundaries can then form and evolve over time, possibly leading to the initiation of subduction and therefore Earth-like plate tectonics. In general, differences in rheology properties of the top planetary layer can result in different types of convection [113,114]. Typically three different convective regimes are distinguished in addition to plate tectonics: stagnant, sluggish/transitional and mobile regimes [109,114]. Stagnant lid convection is characterized by a very viscous lid on top of an actively convecting mantle and forms for viscosity contrasts of at least 4 to 5 orders of magnitude between the mantle and the surface layer, depending on the Rayleigh number and creep mechanism. This is the case for example for Mars, present-day Venus, Mercury, the Moon, and many icy moons of the Solar System. For small viscosity contrasts (1 to 2 orders of magnitude for diffusion creep and 5 orders of magnitude for dislocation creep), convection is close to isoviscous convection, or mobile regime convection, in which there is a continuous exchange of material between interior and surface. A hot and more ductile surface instead of a cold brittle lithosphere as for a stagnant lid might lead to a sluggish, ductile convective regime at the surface characterized by a surface mobilization with very weak material transport (as has been suggested for the early Earth [115–117] and past Venus [118]). Such a situation may occur for exoplanets orbiting in close proximity to their host star, for which the surface is strongly heated by stellar radiation (possibly only on one side if the planet is tidally locked). For even higher heating, the surface of the day-side of close-in planets can be covered entirely by lava, as has been suggested for 55 Cancri

e [119]. Depending on the heat variation from surface to mantle and on the resulting global viscosity contrast, convective motions from the mantle can then reach the surface, characteristic of mobile lid convection.

Several studies have addressed the question of the occurrence of plate tectonics on rocky planets as a function of global observable parameters, with the aim to be able to predict which exoplanets may be more or less prone to having its lid break up into subducting plates. One of the main factors determining the ability to create stress patterns in the lithosphere that can initiate plate boundary formation, is the internal heat (both initial mantle temperatures after planet formation and concentration of radioactive heat sources). Hot mantles lead to vigorous convection in the planet's interior, but fail to create high-enough convective stresses at the base of the lithosphere [e.g. 120]. By comparison, convection in very cold mantles may be too sluggish, leading to a thick, unbreakable lithosphere. There seems to be a sweet spot of mantle heating rates that favour initiation of plate tectonics [e.g. 121], even though the exact temperature range needed in the upper mantle depends on several other factors such as the mass of the planet. The ratio of surface area to mantle volume defines the cooling efficiency of the mantle, and scales approximately with reciprocal planet radius. For a planet like Mars, being one tenth in mass of Earth, the mantle initially cooled very quickly, leading to a thick lithosphere, and plate tectonics would only have been possible in the very beginning of the planet's evolution. For super-Earth planets with masses several times that of the Earth the case is less clear. A smaller surface to volume ratio suggests less efficient cooling of the mantle, a hot interior and hence a decoupling of the low-viscous mantle convection and the lithosphere. The resulting low convective stresses acting at the mantle-lithosphere boundary would lead to a long-term stagnant-lid regime [e.g. 122]. On the other hand, a high Rayleigh number (Equation 5) leads to generally faster motion in the interior and a thin and less strong lithosphere, which may help to trigger local failures of the lithosphere [e.g. 123]. Since the Rayleigh number depends on the cube of the mantle thickness, a larger planet would be expected to have a much higher Rayleigh number than Earth, and therefore plate tectonics initiation should be more likely than on Earth [e.g. 124]. Several studies have investigated this apparent paradox and have led to seemingly contradicting predictions for super-Earth planets, from plate tectonics being less likely than on Earth [e.g. 125], independent of mass [e.g. 126, proposing water to have a stronger effect than mass] Kor10 to being more likely with increasing mass [e.g. 98,127,128]. Noack and Breuer [120] and Stamenkovic and Breuer [129] showed that the apparent disagreement between mass-dependent plate tectonics initiation studies can be solved by combining information on other factors such as initial mantle

energy, concentration of radioactive heat sources, water content in the mantle, and viscosity variations in the mantle.

The dependence of transport properties on temperature and pressure can also affect significantly the heat flow through the mantle and the evolution of the interior mantle temperature of massive exoplanets. Since viscosity is expected to increase strongly with depth for super-Earths planets more massive than Earth, convective velocities strongly decrease with depth and can lead to sluggish convection in the lower mantle above the core-mantle boundary of massive super-Earths [e.g. 44,45], although the deep mantle may also be hot enough for hot rising plumes to form due to the temperature dependence of the viscosity [98]. In contrast to exoplanet studies neglecting the pressure-dependence of the viscosity, which overestimate the cooling efficiency for massive rocky planets, the lower part of the mantle cools slowly, irrespective of whether the cooling rate of the convecting top part of the mantle is determined by plate tectonics or another convection regime like stagnant lid. Stamenkovic et al. [44] suggested that the temperature evolution in the lower part of the mantle becomes approximately independent of the surface regime with increasing planetary mass even though temperatures in the upper mantle depend strongly on it. On the other hand, Karato [130] suggested that an additional mineral phase change in the lower mantle as well as a change from vacancy to interstitial diffusion can lead to a decrease in viscosity by several orders of magnitude for pressures beyond the core-mantle-boundary pressure of Earth. The viscosity contrast between the top and bottom of the mantle then increases for more massive planets than Earth and tends to reduce the surface mobility and to make stagnant-lid convection more likely [131].

Another factor that strongly influences mantle convection is the compressibility of rocks and the resulting effects on mantle thermodynamic and transport properties. Compressibility tends to suppress convection in the lower mantle [e.g. 132] due to 1) a higher density for larger pressures, leading to less buoyancy of the material in the lower mantle, 2) a smaller thermal expansion coefficient, and therefore a smaller thermal buoyancy effect [e.g. 133] leading for example to deceleration of downwellings [e.g. 134], and 3) an increased thermal conductivity and viscosity, leading to more transport of energy by conduction and a stabilized sluggish lower mantle [e.g. 135–137]. The temperature in the lowermost part of the mantle is important since it partly determines the heat flow out of the core and hence the possibility to generate a magnetic dynamo field in the core, if at least partially liquid. The surface convection regime (for example plate tectonics or stagnant lid) on the other hand strongly influences the (evolution of) surface conditions and atmosphere, and plays an important

role in the possibility to have water and hence Earth-like life at the surface (see Section 6).

### 5.3. Outgassing

During the solidification of a magma ocean, massive outgassing of volatiles occurs, called *primary outgassing*. This can lead to a dense primary atmosphere with partial pressures on the order of hundreds of bars for  $\text{H}_2\text{O}$  and  $\text{CO}_2$  [138], resulting in high temperatures at the surface of the planets [139]. Substantial amounts of volatiles remain stored in the rocky mantle during the solidification phase [140], depending on how fast the magma ocean solidifies and on interactions with the steam atmosphere [141]. The pressure of the atmosphere influences the solubility of volatiles in the magma [e.g. 142,143] and therefore how many and which volatiles will finally be stored in the solidified mantle. The solubility varies also with volatile concentration and melt composition. Water and sulfur have a lower solubility than  $\text{CO}_2$ , and reduced species such as  $\text{H}_2$ ,  $\text{CO}$  and  $\text{CH}_4$  have an even lower solubility than water [e.g. 144, and references therein]. Outgassing from the interior can therefore also occur in later evolution stages, called secondary outgassing, if mantle rocks can melt and if the melt rises towards the surface. Rocks can melt for various reasons, for example when they rise adiabatically as a result of convection and their temperature becomes higher than the local melting temperature, which decreases with decreasing pressure. Volatiles like  $\text{H}_2\text{O}$  and  $\text{CO}_2$  preferentially partition into melt, and even for small degrees of melting, large amounts of volatiles exceeding that of the initial rock concentration can accumulate in the melt. When melt is extracted from the silicate matrix, the residual material is depleted in volatiles (see Figure 7). Extrusive volcanism leads to exposure of the melt to the atmosphere at the surface, and outgassing will occur depending on the solubility of the volatiles and the redox state of the melt. Due to the high solubility of water and sulfur in the melt, at atmospheric pressures such as for Venus' surface, outgassing would be dominated by  $\text{CO}$  and  $\text{CO}_2$  [144]. For pressures below a few bar, the composition of gases would change and include also other gases such as  $\text{H}_2\text{O}$ ,  $\text{H}_2$ ,  $\text{SO}_2$  and  $\text{H}_2\text{S}$ . The redox state of the melt defines the gas species that are released into the atmosphere [e.g. 145, from reduced (e.g.  $\text{H}_2$ ,  $\text{CH}_4$ ,  $\text{H}_2\text{S}$ ) to oxidized species (e.g.  $\text{CO}_2$ ,  $\text{SO}_2$ ,  $\text{H}_2\text{O}$ ). Intrusive magma can react with surrounding crustal rocks to form metamorphic mineral phases, and depending on metamorphic reactions can contribute at a later stage to atmospheric degassing [146,147]. For melt to rise towards the surface, its density needs to be smaller than the density of the surrounding rocks. This is generally the case if the local pressure is below about 8–20 GPa

depending on the temperature as well as on the composition of melt and solid rock. For larger pressures, the melt can be denser than the surrounding rocks, since it is often more compressible, and therefore, it sinks deeper into the mantle [103,104,148] where it can recrystallise.

Secondary outgassing strongly depends on the mass, composition, interior structure and dynamics of the planets, and therefore also evolves over geological timescales. With increasing mass or iron to silicon ratio secondary outgassing of greenhouse gases on stagnant-lid planets diminishes [45,149]. Both lead to an increased pressure gradient in the lithosphere, and although the lithosphere would be expected to be thinner for more massive planets due to the increased interior heat budget (since a larger mass correlates with more gravitational and accretional energy during planet formation), the pressure at the bottom of the lithosphere is higher than for an Earth-mass stagnant-lid planet. As a result, partial melting and volcanic activity decrease or even shut-down with increasing mass and iron to silicon ratio [45,105]. Although the initial heat after the magma ocean solidification and the amount of radioactive heat sources in the mantle influence the amount of volcanic activity that can occur on stagnant-lid planets [105], massive super-Earths may not contain atmospheres rich in greenhouse gases. If plate tectonics is active on a planet, mantle material is transported upwards towards the surface at diverging plate boundaries, where pressures are low enough for pressure-release melting to occur. Planets with plate tectonics are therefore expected to have a significant atmosphere, but volatiles can be subducted back into the mantle via for example hydrated rocks and carbonates. While destabilization of minerals with increasing temperature and pressure can lead to a shallow release of volatiles back to the surface, some water and carbon can be transported into the deep mantle [150–152], replenishing the volatile content of the mantle and allowing for long-term outgassing of volatiles, which can then stabilize the atmosphere over long geological time scales.

## 6. Habitability

### 6.1. Habitable conditions for Earth-like life

In everyday life, the words ‘habitability’ and ‘inhabited’ are typically used with respect to humans, as an uninhabited area on Earth does not mean that there is no life there at all, but that there is no human occupation of the land. In science, planetary habitability is typically used in a broader way, meaning that a habitable planet has the right conditions to host any kind of life. In a more restricted sense it most often refers to Earth-like life since that is the only life form for which we understand at least to some

extent the conditions required for life to originate and evolve (even though we still do not know precisely where and when life originated on Earth). The basic requirements for life as we know it are availability of building blocks of life (the CHNOPS elements), efficient energy sources, chemical nutrients, and liquid water as a solvent [153]. On different, more exotic worlds, other solvents like methane or ammonia might make an environment habitable at significantly lower temperatures than on Earth. Life on such exotic worlds would be very different from life as we know it, which would make the detection of such life forms challenging if not impossible. As is usual in exoplanetary studies, we therefore restrict habitability to Earth-like life.

In the Solar System, water is not very abundant at the surface of planetary bodies (in the form of oceans, rivers, rainfall). The largest liquid water reservoirs can be found under an isolating ice shell in the form of a global subsurface ocean in large icy satellites. Life may form under these conditions if chemical nutrients (produced by reactions of water with rocks) are abundant and a sufficiently high energy source is available. These conditions can be met if liquid water is in direct contact with a rocky material, a situation that occurs if no high pressure ice layer separates the water ocean from the rocky layer below [49,154]. The Jupiter moon Europa and Enceladus, a moon of Saturn, are the only two bodies in the Solar System currently thought to have such an interior structure. They are considered to be the most interesting targets for future in-depth investigation of their habitability [155].

## **6.2. Water-rich exoplanets**

Besides Earth-like exoplanets, water-rich planets of sub-Neptune masses with an assumed rocky inner part (possibly differentiated into a silicate layer surrounding a metallic core) and a deep H<sub>2</sub>O (water and/or ice) top layer gained much attention in the past decade. These water-rich planets can, depending on their formation history, either possess a dense, hydrogen-helium atmosphere above the H<sub>2</sub>O layer, such as e.g. Neptune or only have a thin or no atmosphere, like the large icy satellites of the Solar System. The surface of the latter planets may be either ice or water depending on the pressure and temperature conditions. In the H<sub>2</sub>O mantle, as explained above, phase transitions can occur between water and ice and between different high-pressure ice phases depending on the temperature and pressure and the extent of the H<sub>2</sub>O layer. High-pressure ice forms if the H<sub>2</sub>O layer exceeds a critical depth. For a surface temperature of 290 K, the critical depth for the phase transition is at about 160 km for an Earth-mass planet compared to 50 km for a 10 Earth mass planet [e.g. 49,156]. If the H<sub>2</sub>O layer does not exceed that depth, the planet

is considered habitable since then the H<sub>2</sub>O layer is entirely liquid or only the crust is ice and so the water layer is in direct contact with rocks. Subsurface oceans might be the most common place for habitability in the universe.

If high-pressure ice forms in such a planet, a habitable environment may still exist if ice can melt at the bottom of the ice layer above the rocks. Depending on the depth of the ice layer, bottom ice melting due to inefficient heat transport by convection through the high-pressure ice layer and accumulation of water can lead to a possible habitable niche inside such a water-rich planet [49]. Another potentially habitable area would be the liquid water above the high-pressure ice but this would require the water reservoir to be enriched in nutrients for lack of direct contact of the water with rocks. The nutrients can either be delivered to the ocean layer by impacts (which could also deliver organic material), or can be transported by sub-ice-layer water, that has been in contact with the rocky part of the planet, through the convecting high-pressure ice layer [49].

Outside of the Solar System, it will be very difficult to find life in subsurface oceans under an ice shell [157]. For remote detection of life on an exoplanet or -moon, detectable features on the body's surface and atmosphere are needed. For thin ice shells, interaction with the surface might be possible through for example geysers (such as for Enceladus). Although the search for life outside the Solar System usually concentrates on rocky, Earth-like exoplanets with water exposed at the surface, planets with habitats below the surface might be more common and require further study.

### **6.3. Habitable zone**

Although habitable reservoirs can exist below the surface of a planet, the *habitable zone* (HZ) of a star is classically defined as the range in distances of a planet to its host star where water could be liquid at the surface, if the atmosphere contains the right amount of greenhouse gases and clouds [158,159]. An atmosphere is needed in general to allow for water at the surface [154]. The planet must therefore have enough volatiles and be sufficiently massive in order to be able to retain its atmosphere. At the outer boundary of the habitable zone, a sufficiently dense atmosphere is needed to avoid sub-zero temperatures. Clouds can further contribute to increasing the surface temperature, an effect which can also be observed on Earth. For example, after a cloudless night, temperatures are colder at the surface than after a cloudy night, since heat from the surface can radiate to space unhindered. On the other hand, a thick cloud layer hinders heating of the surface by solar radiation during the day. An excess in oxygen in the atmosphere (even though essential for human life) is not a necessary

requirement for habitability, since on Earth it evolved (at least partly) as a result of life [160].

The classical HZ has been defined for Earth, assuming that Earth can compensate variable luminosity (which changes with age) at different orbits around the Sun by variations in the atmosphere – for example by enhanced weathering closer to the Sun or through increased carbon release from carbonate rocks and volcanic outgassing further away from the Sun. We note that a calculation of the limits of the HZ based on the surface temperature that results from an equilibrium between stellar irradiation and re-radiated blackbody radiation from the planet's surface would place Earth outside the habitable zone since the atmosphere increases the mean surface temperature of the Earth by about 30 K. The equilibrium surface temperature can be expressed as

$$T_e = \left( \frac{(1 - \alpha)S_0}{4\sigma} \right)^{\frac{1}{4}}, \quad (6)$$

where  $\alpha$  denotes the surface albedo,  $\sigma = 5.67 \times 10^{-8} \text{ Wm}^{-2}\text{K}^{-4}$  is the Stefan-Boltzmann constant, and  $S_0 = L/(4\pi D^2)$  expresses the incoming stellar flux at a distance  $D$  from the star with luminosity  $L$ . For the Earth with an albedo of  $\alpha = 0.3$ , a distance of approximately  $1.5 \times 10^{11} \text{ m}$  to the Sun, and a solar luminosity of  $L_{\odot} = 3.8 \times 10^{26} \text{ W}$ , the present-day equilibrium surface temperature is 255 K or  $-18 \text{ }^{\circ}\text{C}$ .

The *continuous habitable zone* is defined as the region where the surface temperatures of the planet allow for liquid water during most of the planet's evolution [158]. The inner boundary of the HZ is defined as the distance from the star where water vapour becomes an efficient greenhouse gas and leads to a runaway greenhouse effect as suggested for Venus [161]. The outer boundary of the HZ is set at the maximal distance from the star, where other greenhouse gases such as carbon dioxide and methane are not efficient enough anymore to raise surface temperatures above the freezing point of water, which is called the maximum greenhouse limit. The classical HZ therefore can only be applied for planets that are somewhat similar to Earth, and where the atmosphere is composed of varying amounts of  $\text{N}_2$ ,  $\text{CO}_2$  and other greenhouse gases depending on the distance from the star. For different greenhouse gases (e.g. primordial amount of H-He, methane-rich atmospheres, etc.) the habitable zone would lie at very different distances to the star. For hydrogen-dominated atmospheres, the habitable zone could extend up to about three times the habitable zone width defined for an Earth-like atmosphere [162,163].

The habitable zone depends on the luminosity of the star. With decreasing mass of a main sequence star, the stellar luminosity also decreases and the habitable zone moves closer to the star. Planets around M dwarfs (cool



stars with a mass between 0.075 and 0.5 solar masses) with an orbit of several days may well lie within the habitable zone of that star. This is for example the case for the planet Proxima Cen b, which orbits our closest neighbour star in the Alpha Centauri system [164], and several planets in the densely packed TRAPPIST-1 system [9]. Planets orbiting in the habitable zone of M dwarfs are more easily detectable compared to a planet orbiting the habitable zone around a Sun-like star because of the smaller orbit and smaller central star. M dwarfs are also by far the most common type of stars in our galaxy, which makes them top candidates for exoplanet detection missions. The position of the habitable zone of M dwarfs is more constant in time than for more massive stars. Since the luminosity of a star evolves over time, also the habitable zone slowly shifts. The evolution of the luminosity mainly depends on the mass of the star: low mass stars, such as M-dwarfs, evolve very slowly and can be stable for over hundred billions of years, whereas our Sun, which is a G-type star, has a total main-sequence lifetime of only 10 billions of years. Stars off the main sequence evolve in a more complicated way and often much faster.

Planets that lie in the habitable zone are not necessarily habitable since the classical definition of habitability does not take into account all factors that could affect the planet's potential habitability. The activity of stars, which may be very strong in the beginning of the star's evolution depending on the type of star, can erode a planet's atmosphere and strongly reduce habitability. In particular, planets around M dwarfs probably suffer strong atmospheric erosion due to high EUV exposure and flares (as has been suggested for Proxima Cen b [165]). Also the magnetic field of the star could potentially heat up planets on a close orbit by induction heating to a surface temperature too hot for habitability (as has been proposed for the three close-in planets around TRAPPIST-1 [43]). On the other hand, planets discovered in the HZ might experience a too strong greenhouse effect if atmosphere sinks (erosion of atmosphere to space and formation of volatile-bearing rocks at the surface) are insufficient, turning the planet into a more Venus-like state. Or it could be that more massive planets accreted and kept considerable amounts of H and He gas (primordial atmosphere), leading to surface temperatures above those for liquid water and possibly even to magma oceans at the surface of rocky planets in the HZ.

Planets may also have formed under dry conditions or lost all water from the interior during the solidification stage of the magma ocean. Super-Earths with masses above a couple of Earth masses might not be able to build up an atmosphere dense enough to allow for liquid water at the surface if they are in a stagnant-lid regime [45,105]. These planets might therefore not be habitable, even if detected within the habitable zone, unless they could maintain parts of the primordial or primary

outgassed atmosphere, or they evolved into the plate tectonics state associated with significant outgassing. Another positive role of plate tectonics on the long-term habitability of a planet is the continuous replenishing of the surface with fresh crust, which is needed to take up carbon via seafloor weathering (a crucial factor for the global carbon cycle) and leads to continuous availability of nutrient-rich soils. Plate tectonics can thus help to regulate the atmosphere of a planet over geological time scales via volcanic outgassing and volatile subduction, if sufficient amounts of volatiles have been accreted and survived in or on the planet.

A further strong link between habitability and the planetary interior occurs for planets in the habitable zone of low-mass stars. Since those planets are closer to the star than for a solar-type star, because of their smaller luminosity, they have much more intense tidal interactions. Since the surface temperature approximately scales as  $L/a^2$  (neglecting atmospheric effects), where  $L$  is the luminosity of the star and  $a$  the distance to the planet, the orbital distance of the planet  $a$  for a given surface temperature scales as  $a = \sqrt{L}$ . For the TRAPPIST-1 planetary system, the luminosity of the central star TRAPPIST-1 is  $5 \times 10^{-4}$  that of the Sun Gillon2017, indicating that a planet like the Earth would have about the same surface temperature around TRAPPIST-1 as around the Sun at a distance of about 0.022 AU. This distance corresponds to the distance around which the TRAPPIST-1 planets orbit their star. Tidal friction has a much stronger dependence on distance and results in a much faster despinning rate  $\sim M_{\text{star}}/a^6$  of the planet than for Earth, which is almost a billion times faster than for the Earth. As a result, those planets could be tidally locked to the star in a 1:1 spin-orbit resonance [166,167], in which the planet always shows the same face to the star like our moon does to the Earth. In such a case, the near-side of the planet would experience considerable heating, whereas the far-side would never receive any direct heating from the star. Temperatures at the night side can then only rise above the background temperature of space if an atmosphere regulates the temperature dichotomy, which can lead to extreme weather conditions on the planet [e.g. 168,169].

## 7. Conclusions and perspectives

Although information about the physics and chemistry of exoplanets is very limited at the moment, scientists and the public at large have a desire to know what is the nature of planets outside the Solar System and whether they would be hospitable to life. The question of whether we are ‘alone’ in the galaxy or universe, is a strong driver to study habitability even at this early stage of exoplanetary exploration. The large number of exoplanets

detected has already demonstrated a large diversity in exoplanets and theoretical progress has given some ideas about possible exoplanetary habitats.

The planetary interior plays a crucial role in shaping (sub)surface conditions in which life could thrive. It not only plays a dominant role in creating and sustaining the topography and determining the mineralogical structure of the top layers, but also determines the heat flow from the center to the surface and thereby regulates the occurrence of volcanism, the existence of a global magnetic field, and is the source and a driving force behind the evolution of the atmosphere. Although the data are currently scarce, investigating planetary interiors is essential for an in-depth understanding of exoplanets and for assessing their habitability. It also is the prime method of study for covered areas of habitability such as subsurface oceans.

Planetary interior modellers evaluate possible interior structures and investigate various evolutionary pathways of exoplanets. They are guided by the much more detailed knowledge of the Solar System planets and moons, but also extend what is known about the Solar System to the uncharted exoplanetary world. One of the key problems in those studies is that knowledge about planetary materials is usually only available up to at most the pressure and temperature in the core of the Earth and for compositions that are Earth-like. Several initiatives are being undertaken to extend these studies beyond Earth conditions. For example, thermo-elastic properties of an iron core are studied by *ab initio* calculations up to the highest pressures expected in exoplanets [e.g. 50,170] and by laboratory experiments up to 1.4 TPa [171,172]. Also rocky materials of different compositions are being studied, for example for carbon-rich planets [e.g. 173–175]. In order to advance our understanding of exoplanetary structure and dynamics, further extension of such studies are needed in terms of wider composition, pressure and temperature ranges and more physical quantities.

A main obstacle to understanding planetary interiors and habitability is the absence of compositional data on a large sample of exoplanets. Without data on the composition of planets, the mass-radius diagrams are highly degenerate [e.g. 50,51]. A promising method to obtain compositional information is through spectroscopic observations of the atmospheres of planets. For a habitable planet with water at the surface, the atmospheric composition is not representative of the bulk composition of the planet, although related to it in a complicated way through outgassing, weathering, condensation and possibly subduction in plate-tectonic planets. With increasing surface temperature, the atmospheric composition becomes more representative of the bulk composition [176–178]. Since most observed exoplanets have a much hotter surface than the Earth, and a significant fraction has an expected surface temperature above 1000 K, bulk

compositional data can be obtained for metals, refractory elements and volatiles from spectroscopic observations of their atmosphere. Such observations will be done on a large sample of planets (> 1000) with the ground-based ELT (first light targeted for 2024), and with space missions like the JWST (James Webb Space Telescope, to be launched in 2020) and the recently selected ARIEL mission of ESA [179] (due for launch in 2028), and will provide a much refined view on interiors of hot exoplanets, that will also help in better understanding habitable exoplanets.

Additional information on the interior can also be derived through interpretation of other observational data (in)directly influenced by the interior. For example, rotational characteristics of exoplanets can depend significantly on tidal dissipation inside the planets. The rotation of an exoplanet has only been determined for the first time in 2014 for the gas giant  $\beta$  Pictoris b [180] but future rotation data for rocky or water-rich exoplanets could be used to indirectly infer the presence of regions of high dissipation such as in Io or liquid layers as in subsurface oceans. The shape of exoplanets, measurable from the light curves during transits, can also be used to reduce the degeneracy in mass-radius diagrams since deviations from spherical symmetry, caused by rotation and tides, depend on the interior. This technique is particularly promising in assessing the iron to rock ratio in super-Earth exoplanets [181]. Further advances are also expected from studies of the elemental abundances of the host stars, which will shed more light on the composition of their orbiting planets, for example on the abundance of long-lived radio-active elements, which play a crucial role in the thermal evolution of planets.

In addition, the statistics of exoplanetary data will see a major extension with upcoming ground-based and space mission projects, such as TESS (launched on 18 April 2018), CHEOPS (to be launched at the end of 2018), and PLATO (to be launched in 2026). This will guide the theoretical studies and will lead to a more refined and complete classification of planetary diversity. Together, the observational data and further advances in theoretical planetary modelling and in the behaviour of planetary material under exoplanetary conditions, will ascertain major progress towards a deeper understanding of planetary interiors and evolution, and a better assessment of their habitability, even if detection of life itself beyond Earth may not be achieved in the near future.

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