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Interior Controls on the Habitability of Rocky Planets

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No matter how fascinating and exotic other terrestrial planets are revealed to be, nothing generates more excitement than announcements regarding their habitability. From the observation of Mars to present-day efforts toward Venus and the characterization of exoplanets, the search for life, or at least environments that could accommodate life, has been a major drive for space exploration. So far, we have found no other unquestionably habitable world besides Earth. The conditions of the habitability of terrestrial planets have proved elusive, as surface conditions depend on the complex interplay of many processes throughout the evolution of a planet. Here, we review how the interior of a rocky planet can drive the evolution of its surface conditions and atmosphere. Instead of listing criteria assumed to be critical for life, we discuss how the bulk-silicate planet can affect the onset, continuation, and cessation of habitability. We then consider how it can be observed, and current efforts toward this end.

Introduction

Investigations into habitability have provided ample opportunity for not only reflection as to its nature but also planetary evolution processes in general. Agreeing on a definition for habitability has proved troublesome. Rather than providing a definition, we consider that the concept can be distilled to a simple question: Can life develop there? Unfortunately, the notion of life is heavily biased toward life as we know it on Earth, and even the question of "development" remains vague: Is survival enough? Does it include its proliferation instead? Does it cover the emergence of life? Would the requirements not change with time and because of life itself? The interested reader could consult the review by Cockell et al. [1] for a discussion of the definition of the term.

It is nonetheless possible to draw a list of requirements that any living organism we can imagine would need to remain active. Those would include (a) a solvent (water, on Earth), (b) energy, and (c) the presence of various elements and nutrients in sufficient abundances and forms to facilitate (a) and (b).

It has been argued that energy was likely easily accessible in general, from solar radiation, redox potential, or other processes, and that life as we know it makes use of some of the most common building blocks (elements) in the universe, the CHONPS (carbon–hydrogen–oxygen–nitrogen–phosphorus–sulfur). Even minor elements that are used by life, such as Fe or Mg, for instance, are relatively abundant on the surface of Earth (even if those may vary from one specific organism to the other). Finally, the solvent common to known life, water, is one of the most abundant molecules in the Solar System.

Indeed, spectroscopic surveys show that compositions of stars in the solar neighborhood are similar to that of our Sun [2]. However, despite the ubiquity of the right elemental building blocks, the universe does not appear to be teeming with life. This indicates the need for the appropriate physicochemical conditions to arise on planetary surfaces. For example, water would need to be liquid, which imposes further constraints on the extant pressure and temperature conditions.

There are also important differences between conditions required for the emergence of life and its continuation: No credible model for abiogenesis has been proposed, and we are still unable to produce new life through laboratory experiments and have not recorded spontaneous life formation [3,4].

Many specific characteristics have been proposed as essential to planetary habitability (see [1]). Some are related to a planet's galactic location, its star and orbit [5], the pattern of the other planets in the system, or the presence of a large moon [6]. Other are related to the planet's properties like its composition [7], inherited from its origins (see the "The Onset of Habitability" section), its size/mass (see the "The Role of the Convection Regime" section), or processes (such as plate tectonics, for example [8]).

However, it has been suggested that piling up these criteria will inherently result in a rare-Earth-type situation [9] that only an infinitesimal fraction of terrestrial planets could fulfil [10]. In short, precise lists describe only Earth. While it is possible that Earth could be the template for all habitable worlds and that those are incredibly rare, limiting our approach leaves gaps in our understanding of life in general.

In the end, on the basis of the above, if a single criterion is to be kept, it should be the presence of liquid water. Disregarding the characteristics of individual organisms, in terms of biology, liquid water is deemed necessary both for the emergence and persistence of life (e.g., [11]) and therefore roughly demarcates a range of surface conditions amenable to the occurrence of life. Investigating how a planet can reach and maintain conditions compatible with the presence of liquid water remains the best general purpose criterion we can apply.

This idea was the seed for the development of the concept of habitable zone (see [12,13], for example, but references to this concept date back to at least the 19th century). The circumstellar (abiotic) habitable zone (HZ) [14] is defined as the shell of space around a star where liquid water can be maintained at the surface of a planet. Its inner boundary is set by the maximum energy flux that can be received in the presence of liquid water (see the "The Cessation of Habitability" section). Its outer boundary is set where the maximum greenhouse can no longer sustain temperatures high enough for liquid water [15]. Most depictions of the HZ assume an Earth-like planet with an Earth-like atmosphere, highlighting the importance of atmospheric composition.

A mix of N_2, CO_2, and H_2O [16] is often considered, but other gases would contribute, like CH_4 or H_2 (e.g., [17]). The HZ also varies with the type of star considered and its age, as its luminosity changes throughout its evolution (e.g., [18,19]).

This simple definition has historically been used to select targets that are deemed likely to be habitable (see, e.g., [20]).

A more complete definition would consider the whole planet as a dynamic system, possibly including life itself. It remains uncertain how such extended definitions of HZs could be translated into practical parameters or observables, but future observation and characterization of exoplanets will help refine this discussion.

The presence of liquid water implies sufficiently clement conditions that H_2O (liquid) is more stable than its gaseous equivalent, water vapor. Therefore, the habitability of a terrestrial planet depends on the nature of its atmosphere. Hence, processes that affect the volatile inventory (such as the accretion history or atmospheric escape) or the atmosphere response to solar radiation (radiative transfer, the greenhouse effect or atmosphere dynamics) influence planetary habitability. However, as one of the drivers of long-term planetary evolution, the state of the interior of terrestrial planets is critical to the origin and evolution of the atmosphere and surface conditions. The dynamics of the mantles of planets provides a link between the atmosphere/surface and the deep interior, which extends down to the core [21,22]. Mantle thermal history, dynamics, and eventually melting cause volcanism and volatile delivery to the atmosphere. Interior dynamics also play a part in the regulation of the climate on Earth, by recycling surface material and volatiles into the interior. This cycle is deeply intertwined with outgassing, weathering, and subduction. On Earth, it notably affects the CO_2 concentration in the atmosphere and is named long-term carbon cycle. Likewise, the thermal evolution of the core governs the generation of a magnetic field that affects atmospheric escape processes and long-term atmospheric evolution.

A comprehensive picture of planetary evolution requires consideration of the planet as a complex system that depends on the efficiency of numerous feedback mechanisms. Here, we focus on aspects relative to the interior of rocky planets and only include succinct descriptions of the interactions between atmosphere-based mechanisms and the interior of planets for clarity. The present review focuses on the relationship between habitability and the silicate planet. Initially and to set the boundary conditions from which a given rocky planet is thought to have evolved, the molten state of the mantle is considered, before transitioning to a solid state and its long-term interactions with the surface and atmosphere of the planet.

Habitability has been described in the context of targeted studies of specific planets like Mars [23] or Venus [24] or in terms of a descriptive approach detailing each mechanism that can affect habitability and the atmosphere (e.g., [22]). Here, we instead draw from the core science questions that remain unsolved, summarize the current state of our knowledge, and discuss how future work can build upon this foundation. We have identified 3 broad questions that are the pillars of ongoing investigations into the habitability of terrestrial planets in the Solar System and beyond:

- How do planets become habitable and to what degree is it influenced by their accretion history, bulk composition, and initial conditions?
- How do the interior dynamics of a planet and its mantle convection regime affect habitability?
- What can cause habitability to end or push a planet out of the regulating cycles that maintain habitable surface conditions?

The onset of Habitability

Volatile budgets

Both planetesimal and pebble accretion imply a considerable degree of stochasticity as to the volatile element budgets of growing terrestrial planets [25,26]. Planets accumulate material from a wide range of heliocentric feeding zones [27], regardless of whether condensed material was initially confined to an annulus [28] or ring [29,30] or whether it resulted from inward-drifting pebbles [31]. This stochasticity leads to a gradual decline in elemental abundance with increasing volatility (decreasing condensation temperature; the temperature at which half of the mass of a given element condenses from a gas of solar nebular composition, at a nominal pressure of 10^{-4} bar) in the terrestrial planets [26]. Elements that were sequestered into the core during its formation (siderophile elements) show depletions in the mantle with respect to the "volatility trend" defined by silicate-loving (lithophile) elements (e.g., [32]). Therefore, although the total budgets of siderophile elements in the terrestrial planets are higher than those observed in their mantles, they are inaccessible to geochemical scrutiny and do not play a direct role in determining habitability conditions at the surface of the planet (although they may indirectly contribute, such as in the generation of a geodynamo [33]).

As such, the bulk planetary abundances of key atmosphere-forming moderately volatile elements, of which S and P are part, can be estimated from those of lithophile elements of similar volatility. Conversely, highly volatile elements, notably C, H, and N, due to their low abundances in planetary mantles with respect to the nebular gas or chondritic meteorites (e.g., [34]), are likely to have been delivered irregularly and near the tail end of accretion (e.g., [25,35]). As a result, their provenance is likely distinct from that of major, planet-forming elements (see [36,37]), such that predicting their abundances in the terrestrial planets is fraught with uncertainty.

Through analysis of peridotitic rocks that comprise Earth's mantle, together with inversion of the compositions of products of mantle melting in the ocean basins, the concentrations of volatile elements in the convecting mantle of Earth can be determined (cf. [33,36,37]). The mass of oxygen is overwhelmingly contained in Earth's mantle, bound in silicates and oxides, and can be computed by stoichiometry to better than 2% relative. The moderately volatile elements, P and S, also reside chiefly in the mantle and have abundances that can be determined to a precision of ~20% relative [38,39].

H, C, and N are presumed to exist as a particular species (H_2O, CO_2, and N_2) during mantle melting, and their abundances in basalts are normalized to those of a refractory lithophile element (H_2O/Ce CO_2/Ba or CO_2/Nb) or failing that another volatile element (Ar for N_2) that behaves in a similar manner (e.g., [40]), with uncertainties of 50 % relative. Furthermore, substantial fractions of the budgets of H, C, and N are present on the surface (atmosphere + oceans + crust) [41] of Earth, known to within 20% relative (see [40] and references therein). Together, the abundances of H, C, and N in the bulk silicate Earth (BSE) can be computed within a factor of ~2. With these caveats in mind, the depletion factors of the 6 key elements, CHONPS, in the BSE relative to the solar- and CI-chondritic (Ivuna-type carbonaceous chondrite) compositions are found in Table.

A striking observation is that no more than ~0.4 parts per million (ppm) (by weight; ppmw) of solar nebula matter remains in Earth based on the Al-normalized abundance of H (Table). Of course, this estimate ignores any amount of H potentially dissolved in the core. Assuming a partition coefficient between core and mantle, $D^{H}_{core/mantle} = 10$ [42], then the 120 ppmw of H in the mantle would give 0.12 wt% in the core, consistent with geophysical constraints [43]. Even if $D^{H}_{core/mantle} = 100$ (as permitted by the computations of [42]) and 1.2 wt% of H were to be dissolved in the core, the depletion factor of H relative to Al and the solar nebula (Table) would be ~10^{-5}, lower than or equivalent to, respectively, the solar- and Al-normalized abundances of C or N, even without accounting for the siderophile behavior of the latter pair (e.g., [44–46]).

Therefore, it seems that, compared to the Sun, Earth (or its building blocks) underwent preferential depletion of H relative to similarly volatile elements, C and N, with the noble gases having been depleted by an even greater factor than H [47,48]. Indeed, Aston [47] recognized that "Earth has only 1 millionth part of its proper quota of inert gases", with modern estimates [35,49] ranging from one part in 10 billion (Ne) to one millionth (Xe). Hence, Earth has retained but a minuscule fraction of a solar component (i.e., from the nebular gas).

The origin of the small fraction of volatiles has been reconciled with either (a) dissolution and subsequent loss of a component of the nebular gas (e.g., [50]) or (b) predominantly dry accretion of Earth with a small fraction of volatile-rich, chondritic, and/or cometary material (e.g., [25]). Isotope ratios among volatile elements offer a complementary way to distinguish between these plausible sources (see [51], for a review). The isotopic compositions of noble gases (except He and Ne), H, C, and N in the BSE overlap with those of chondrites [37,52,53], which, themselves, are distinct from that of the solar nebula (as inferred from the Sun and the atmospheres of the gas giants).

Models of dissolution of the solar nebula can account for the observed abundance of H relative to the noble gases [54,55]. Sharp and Olson [54] highlight the plurality of sources that could have been accreted to Earth to satisfy both elemental and isotopic constraints, with N and heavy noble gas abundances being consistent with chondritic and cometary sources, respectively, whereas H and the light noble gases permit a greater fraction of nebular material. However, such a process would have resulted in an overabundance of H relative to the BSE (Table) and an H isotope composition far too light (D/H of ~30×10^{-6}), necessitating isotopic fractionation associated with its subsequent loss via atmospheric escape. These models have, so far, not considered carbon in their treatment, making it unclear whether dissolution of the nebular gas is capable of producing a match to the hydrogen–carbon–nitrogen abundances of the BSE.

On the other hand, to first order, the relative abundances of volatile elements bear a strong resemblance to those in chondrites (Fig. 1) [37], suggesting that the present-day budgets of these elements were brought by such materials [53,56], although Ne isotopes permit a small fraction of its budget in the BSE to be solar in origin [54,57,58]. However, it should be kept in mind that this is relative to the 10^{-7} that remains of the Ne budget of the solar nebula.

The precise nature of the chondritic material that could have delivered these volatiles, and, what fraction of their total budgets

Table. The abundances of the life-essential volatile elements, C, H, O, N, P, and S, in the BSE, CI chondrites, and the Sun. All concentrations expressed in weight fraction. Uncertainties for C, H, and N in the BSE are of the order of a factor of ~2, while P and S are ±20%, with O of ±2%. Data for the BSE are from [35,38,40], for CI from [38], and for the Sun from [304].

	C	H	O	N	P	S
BSE	1.4×10^{-4}	1.2×10^{-4}	0.4433	2.8×10^{-6}	8.7×10^{-5}	2.0×10^{-4}
CI chondrite	0.0348	0.0197	0.459	2.95×10^{-3}	9.85×10^{-4}	0.0535
Sun	2.17×10^{-3}	0.738	6.34×10^{-3}	7.48×10^{-4}	6.44×10^{-6}	3.42×10^{-4}
Depletion (/CI, Al)	0.0014	0.0022	0.341	3.3×10^{-4}	0.031	0.0013
Depletion (/Sun, Al)	1.56×10^{-4}	3.93×10^{-7}	0.169	9.04×10^{-6}	0.033	0.0014

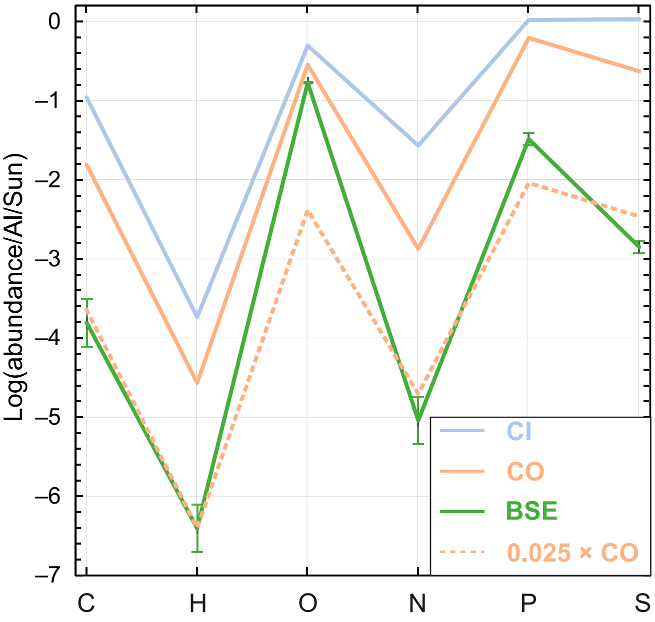

Fig. 1. The abundances of the life-essential volatile elements, CHONPS, relative to the solar abundances [304] normalized by Al in CI chondrites (teal line) [38], CO chondrites (orange line) [52], and the BSE (green line) [35,38,40]. The dashed orange line represents 2.5 % of the CO abundances.

were contributed by such material, remains under scrutiny. Assuming that the D/H and N isotopic compositions of the BSE reflect those of their sources, Earth's H and N (and, by extension, C) budgets were thought to have been carried by CI- or CM-type (Murchison-type) carbonaceous chondrites (e.g., [52,59]). This interpretation has been called into question on the grounds that Earth is known to share an isotopic heritage with enstatite chondrites for the major rock-forming elements (e.g., [36]). Consequently, Piani et al. [53] postulated that enstatite-chondrite-like material could have delivered much of the H budget of Earth. A major caveat is that D/H and N isotope ratios could be fractionated in a mass-dependent manner [45,54] and thus record not only the provenance of these elements but also mass transfer processes (such as atmospheric escape) that could have modified their terrestrial inventories thereafter.

Figure 1 indicates that the addition of 2.5% of CO-chondritic material by mass to an otherwise volatile-free Earth would reproduce the present-day H, C, and N budgets of the BSE, within uncertainty. Equally, 0.7% of CI-like material, consistent with the mass of the late veneer (cf. [25]), would satisfy the H budget of Earth but would deliver 6 times too much N. However, the observation that H, C, and N are depleted relative to the noble gases (normalized to CI chondrites) [49] suggests that a straightforward chondritic contribution is too simplistic and that these elements preserve some vestige of the main accretion phase. Therefore, in the event that chondritic material delivered all of the H, C, and N, this would have represented, at most, ~10% to 25% and 2.5% of the P and O budgets of the BSE, respectively. Hence, P and O must have delivered during the main accretion phase of Earth and hence from predominantly noncarbonaceous material (e.g., [36]). The addition of 0.7% of CI-like or 2.5% of CO-like material would exceed the S budget of the BSE by 1.5 to 2.5×. However, S, together with Se and Te, was likely depleted by the extraction of a small amount of sulfide-rich melt from the mantle after the cessation of core formation but before the accretion of the late veneer [39,60].

In sum, the bulk volatile inventory of terrestrial planets, taking Earth as an example, is low (≤0.05% by mass). This observation, together with the chondritic relative abundances and isotope ratios of major volatiles, indicates chondritic material (although of uncertain origin) as the major source of H, C, and N in the BSE. Their isotopic and elemental compositions are more difficult to reconcile with capture and dissolution of the nebular gas, without appealing to additional processes (i.e., atmospheric escape), and would require further investigation. Instead, it points toward accretion via a mixture of largely volatile-depleted bodies with smaller amounts carbonaceous chondrites (though enstatite chondrites remain feasible [53]), representing ≥30% of the present-day BSE budgets of H, C, and N (see also [61]). Budgets of the less volatile P (≤25% carbonaceous chondrite material) and O (≤2% carbonaceous chondrite material) were brought predominantly by sources other than carbonaceous chondrites, whereas S likely reflects the composition of the late veneer.

Volatile distribution

The constitution of an atmosphere, during a magma ocean epoch, is dictated by the abundances of elements available in planet-building material (see the "Volatile budgets" section), the intensive variables, namely, temperature and pressure that determine the speciation of gaseous molecules and the solubility relations these gaseous species have with their dissolved counterparts in silicate liquids. Because the masses of atmospheres around terrestrial planets are small relative to their interiors, the assumption is frequently made that the interior imposes the state of the atmosphere. As such, the partial pressure (p) of given species, i, in the atmosphere is:

$$p_i = X\frac{gM^a}{4\pi r^2}, \tag{1}$$

where g is the acceleration due to gravity, M^a is the mass of the atmosphere, X is the mole fraction of element i, and r is the planetary radius. To determine the value of M_i^a requires knowledge of the distribution of a given gas species in the magma ocean using a solubility law. A solubility law relates the fugacity of species i, f_i, (or partial pressure, given an ideal gas expected to typify most secondary atmospheres around terrestrial planets) to the mole fraction of the relevant species dissolved in a condensed phase, here, a silicate liquid;

$$X_i = \alpha f_i^{\beta}, \tag{2}$$

where α is a Henrian constant and β is the stoichiometric coefficient. The relationship between the fraction of mass of species i in the atmosphere (M_i^a) relative to that in the mantle (M_i^m), assuming that the sum of the two is equivalent to the total mass, $M_i^T = M_i^a + M_i^m$, and that $f_i = p_i$ with $\beta = 1$, is given by:

$$\frac{M_i^a}{M_i^m} = \frac{4\pi r^2}{gM_T\alpha}, \tag{3}$$

where M_T is the total mass of the planet. Taking a generic chemical reaction and setting $\alpha = 1{,}000$ ppm/bar (in SI units, $10^{-8}\ \mathrm{Pa}^{-1}$) and $\beta = 1$, the distribution of a volatile species, comprising

1,000 ppm of the planet, between the mantle and atmosphere is determined as a function of planetary mass, M_T, where g is given at known r by solving the Adams–Williamson equation of state (see [59,60]).

Figure 2 shows that Eq. 3 predicts that the quantity M_i^a/M_i^m decreases as planetary mass (proportional to g) increases. That is, as shown in [60], small, rocky bodies will have a greater fraction of a given volatile element present in the atmosphere relative to the molten interior as compared to larger planets. Despite the decreasing relative fraction of a given volatile element in the atmosphere, its partial pressure (or fugacity) in the atmosphere increases asymptotically to a maximum value given by Eq. 2; $f_i = X_i/\alpha$. Consequently, mass is a key property that dictates whether the planet is capable of harbouring an atmosphere.

Although some species, notably N_2 [at high oxygen fugacity (fO_2)] [62], CO_2 [63], H_2O (at high fH_2O) [64], and H_2 (at high fH_2) [65] dissolve in silicate melts with $\beta \sim 1$, others have more complicated dissolution stoichiometries. At low fH_2O (≤1,000 bar), water (and possibly H_2 at low fH_2) dissolves into silicate melts with $\beta = 0.5$ [66,67]. At low fO_2, nitrogen dissolves in silicate melts as nitride, N^{3-}, a species that is more soluble than the diatomic molecule [62,68]. The solubility of sulfur has long been known to be redox-sensitive [69–71], existing in silicate liquids as S^{2-} at low fO_2 and SO_4^{2-} at high fO_2.

In this context, the fO_2 holds special significance, as, unlike the other CHONPS elements, O is invariably in excess in the mantle of the planet (and in excess relative to the other major volatile elements; Table) with respect to the atmosphere. Consequently, at equilibrium, the fO_2 of the atmosphere is equivalent to that in the mantle, given by:

$$FeO(l) + \frac{1}{4}O_2(g) = FeO_{1.5}(l), \tag{4}$$

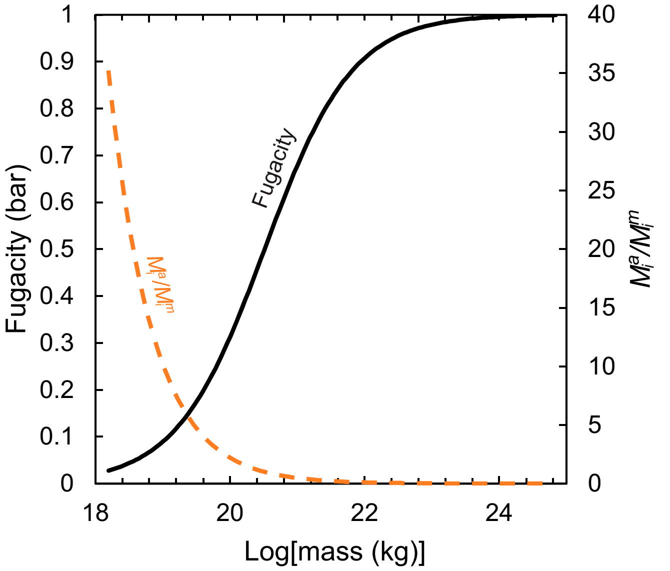

Fig. 2. The distribution of a fictive element, i, with $\alpha = 1000$ ppm/bar and $\beta = 1$ (cf. Eq. 2) between the molten interior (m) and atmosphere (a), expressed in terms of its mass M_i and fugacity (in bar) as a function of the $\log_{10}$ of the planetary mass in kilograms. The planet is assumed to be Earth-like, in which a silicate mantle comprises 70% of its mass, with the remaining 30% being made up of an Fe-rich core.

Sossi et al. [72] exploited this property to determine the speciation and mass of an atmosphere in equilibrium with a magma ocean of peridotitic composition; $Fe^{3+}O_{1.5}/(Fe^{3+}O_{1.5} + Fe^{2+}O) = 0.037$ (that of the present-day Earth's mantle) would set the fO_2 of the atmosphere in equilibrium with the magma to ~IW (IW is the fO_2 at the iron-wüstite buffer; FeO = Fe + $1/2O_2$) (see also [70]). At 2,173 K, the temperature at which peridotite is fully molten at 1-bar pressure, this corresponds to ~10^{-6} bar of O_2. Its importance lies in fixing the relative fugacities of atmospheric species that can be related by fO_2 alone, such as:

$$H_2(g) + \frac{1}{2}O_2(g) = H_2O(g), \tag{5A}$$

$$CO(g) + \frac{1}{2}O_2(g) = CO_2(g). \tag{5B}$$

The CO/CO_2 and H_2/H_2O ratios of the atmosphere, as well as their total pressures, are key in dictating the physical and chemical stability of atmospheres around rocky planets (e.g., [73]). The H_2/H_2O ratio, in particular, influences the fugacities of CH_4 and NH_3, both of which are instrumental in the production of amino acids by spark discharge in the presence of liquid H_2O, as investigated in the Miller–Urey experiment [74]. Their homogeneous gas phase reactions can be written:

$$CO(g) + 3H_2(g) = CH_4(g) + H_2O(g), \tag{6A}$$

$$N_2(g) + 3H_2(g) = 2NH_3(g). \tag{6B}$$

The prevalence of such gases depends on (a) the absolute abundances of H, C, and N in the atmosphere and (b) the atmospheric H_2/H_2O ratio, which is, in turn, governed by the temperature, pressure, and fO_2. To investigate the effect of fO_2 on the chemistry of atmospheres produced on early Earth in equilibrium with a magma ocean, the BSE abundances of the CHONPS elements (Table) are used together with the relative fugacities of stable gas species given by the equilibrium constants [75] of their homogeneous gas phase reactions (e.g., Eqs. 5A, 5B, 6A, and 6B) and their solubilities in silicate liquids (see [76] and references therein). The partial pressures (fugacities) of the major atmosphere-forming species at 2,173 K at 3 different oxygen fugacities (expressed relative to the IW buffer) [77] in equilibrium with an Earth's mantle-sized magma ocean are shown in Fig. 3.

Under reducing conditions (ΔIW-3), early Earth's atmosphere would have had subequal amounts of CO and H_2 (Fig. 3) (cf. [72]), while CH_4 and NH_3 are present at 0.35 and 0.013 bar, respectively. By contrast, S is essentially absent, with the most abundant S-bearing species, $H_2S = 2 \times 10^{-5}$ bar. The total pressure at the IW buffer defines a minimum, ~120 bar, as, at lower fO_2, CO is joined by H_2 and, at high fO_2, CO is replaced by CO_2 and substantial quantities of SO_2 (~80 bar, as well as ~1 bar each of SO and S_2) are present in the atmosphere, a result emphasized by [78]. This reflects the low solubility of SO_4^{2-} in silicate melt at high fO_2 (cf. [79]). Water is not a major component in any atmospheric type as its high solubility ensures that it remains dissolved in the silicate liquid (see also [66,76]), with the implication being that the ~270-bar equivalent mass

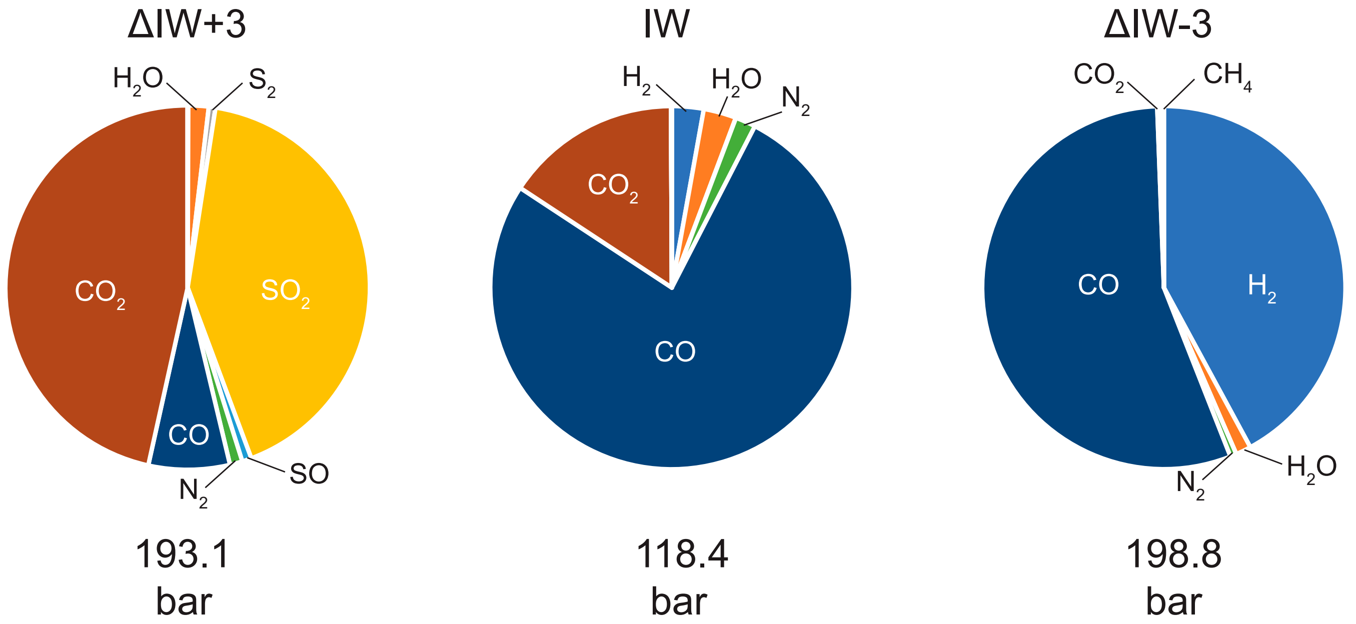

Fig. 3. Equilibrium speciation of atmospheres at 2,173 K formed around an Earth-sized planet in equilibrium with a magma ocean of mass equal to that in Earth's mantle (4.2×10^{24} kg) in the system CHONPS with abundances for the BSE as per Table. The atmospheres are produced given solubility laws for CO [305], CO_2 [63], H_2O [66], H_2 [65], N_2 [62], SO_2 [71], and S_2 [79] and their equilibrium gas speciation solved at 3 oxygen fugacities relative to the IW buffer, +3, 0, and −3. Total pressure in the atmosphere is shown below the respective pie chart.

of H_2O present in the oceans today is a product of the long-term evolution of Earth (see the "The Role of the Convection Regime" section), starting with the solidification of the magma ocean [76,80,81]. That is, owing to the lower solubility of H_2O in solid phases relative to silicate liquid, crystallization tends to expel water from the magma ocean into the atmosphere. However, the degree to which H_2O degasses depends on the mode of crystallization, where convective lock-up of the magma ocean when the melt fraction is ~0.3 prevents further outgassing of water, whereas efficient crystal settling leads to a surface magma ocean and allows water to escape [77]. The sum of partial pressures of phosphorus-bearing species are low, ranging from 2×10^{-3} bar at ΔIW-3 (mainly P, PO_2, PO, and P_2) to 3×10^{-6} bar at ΔIW+3 (chiefly PO_2). Their low contribution results from the nonideality of P_2O_5 in silicate melts ($\gamma P_2O_5 \sim 10^{-10}$) [78,80] and the P abundance in the BSE (87 ppmw; Table).

The relative abundances of gaseous species (Fig. 3) change markedly as the atmospheres cool and components condense and are extracted from the gas phase [72,82]. Notably, CH_4 and NH_3 become increasingly stable down temperature at the expense of H_2 and CO. However, the magma ocean stage is likely to supply the initial budgets of elements to the surface that may then become available to prebiotic synthesis pathways. Greater fractions of S and N are present in more oxidising atmospheres, whereas P- and H-bearing species are stabilized in the gas phase at reducing conditions, while C remains volatile throughout the fO_2 range explored. As such, essential prebiotic molecules, such as HCN, can only form when the partial pressures of H_2, CO, and N_2 are high enough, with its presence being favored at reducing conditions, together with CH_4 and NH_3. This realization led to the notion that transient impact events on early Earth could have locally produced reducing atmospheres [83–86]. Nevertheless, whether these gases can contribute to prebiotic reactions hinges upon the photochemical stability of these species as atmospheres around rocky planets cool over time.

The Role of the Convection Regime

Convection regimes on earth and other planets

Following accretion, rocky planets cool down and crystallize from a magma ocean, leading to the onset of solid-state mantle convection. Different interior dynamics give rise to distinct surface expressions, including habitability. Mantle convection and its surface expressions have a profound impact in the thermal and compositional evolution of a planet, as mantle dynamics link various layers of the planet and drive evolution. In particular, it effectively links the deep interior (down to the planetary core) with the surface/atmosphere through heat transfer and the effect of the magnetic field (see the "Magnetic field generation: An illustration of the far-reaching effects of mantle dynamics" section). Other main expressions of this link include volcanism and the resulting outgassing of volatile species into the atmosphere that, in turn, affects atmospheric mass and surface conditions through the greenhouse effect. In conjunction with other processes (for example, photochemistry and escape), they contribute to the evolution of planetary surface conditions. Together, they effectively produce a coupled system with complex feedback processes, which we are only starting to grasp (i.e., [21,22,87–92]). Here, we focus on the internal component of the system.

Various convective or tectonic regimes have been proposed to exist for different interior states (see Fig. 4), aligning with distinctive observed features. Notably, Earth presently exhibits plate tectonics [93–96]: its outermost rigid shell, the lithosphere, is fragmented into several sizable tectonic plates, gradually shifting atop a mechanically weaker layer called the asthenosphere (part of the upper mantle). Along convergent

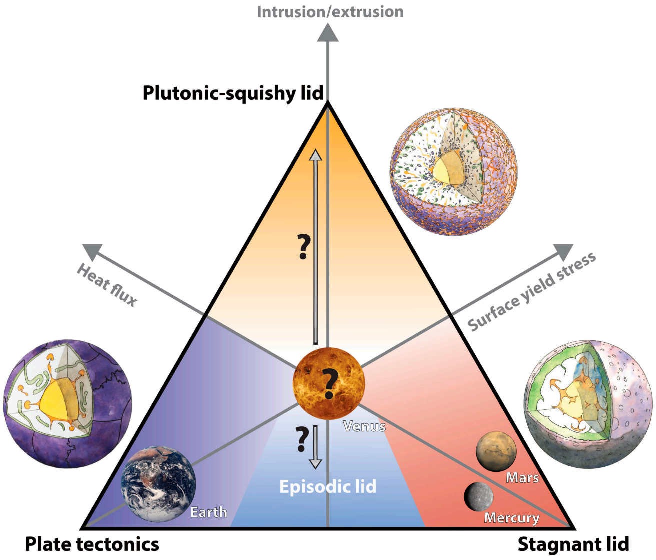

Fig. 4. Schematic depiction of the primary convection regimes observed or theorized on terrestrial planets in the Solar System. The 3 end-members, in this representation, are plate tectonics,stagnant lid, and plutonic-squishy lid. Episodic lid is an intermediate regime between plate tectonics and stagnant lid. While the current regimes of Earth, Mercury, and Mars are well defined, Venus is still uncertain and could range from episodic lid to plutonic-squishy lid or may have transitioned recently (a few hundreds of million years) into stagnant lid. Mercury is in a later stage of stagnant lid compared to Mars because it has cooled down significantly faster. The state of the convection regimes of the planets should not be considered static. Instead, it evolves with time, and the figure can be interpreted as a snapshot of their present-day state. The axis are not fully independent and should be thought of as possible observables. The convection regime figures are adapted from [306].

boundaries, at subduction zones, cold and dense oceanic plates sink into the mantle. This subducted material is balanced by the generation of new oceanic crust along divergent margins and the growth of the lithosphere due to heat diffusion. These dynamics are intrinsically related with mantle convection, i.e., the gradual, creeping motion of Earth's solid mantle.

Advances in numerical modeling have significantly enhanced our understanding of plate tectonics while also revealing other tectonic regimes that can explain observations from other Solar System bodies. These alternative regimes include the stagnant-lid (as observed on Mars, Mercury, or the Moon), where a planet is covered by a single plate [97–99], the episodic-lid regime, characterized by a predominantly motionless lithosphere occasionally undergoing overturning into the mantle [89,100–104], the ridge-only regime, where a large ridge emerges, slightly compressing the surrounding lithosphere without forming subduction zones [105,106], and the plutonic-squishy lid regime, distinguished by a thin lithosphere and several small, strong plates or blocks separated by warm and weak regions generated by plutonism [107,108]. Venus is currently believed to have either an episodic or a plutonic-squishy lid yet the exact nature of the tectonic regime on Venus remains mysterious [22,109].

On Earth, substantial evidence supports the notion that plate tectonics played a pivotal—perhaps indispensable—role in the genesis and development of its atmosphere, oceans, continents, and life (e.g., [110–114]). Increasing evidence indicates that plate tectonics, as a global tectonomagmatic system, has 2 significant impacts: (a) It governs the cycling of essential nutrients between various reservoirs such as the geosphere, oceans, and atmosphere through diverse tectonic, magmatic, and surface processes [115], and (b) the gradual movement of tectonic plates, along with associated topographic alterations and consistent moderate magmatic activity, shaped and fostered biological evolution (e.g., [115–117]).

It is worth pointing out that there are indications that the inception of life might have occurred on an Earth devoid of plate tectonics. This proposition stems from zircon palaeointensity data, which indicates unvarying latitudes for the samples dating back from about 3.9 billion years (Ga) ago to about 3.3 Gyr ago [118], implying that no large-scale surface motion occurred at the time on Earth (contrast for example with [119] and references therein). The earliest indications of life on Earth are older than 3.7 Gyr (e.g., [120]).

Despite its significance on Earth, plate tectonics has not been observed on any other planet. Over the last few years, there has been a notable increase in the discovery of rocky exoplanets, particularly the ones known as super-Earths, which are large terrestrial planets with masses up to 10 times that of Earth, roughly up to twice its diameter. Most of studies indicate that these planets should experience a mobile-lid regime or plate tectonics [121–124]. Nevertheless, some suggest the contrary [125,126]. Therefore, it is also worth contemplating whether life can emerge and flourish on planets with different tectonic regimes.

The interior of terrestrial planets and outgassing

The study of how the interiors of terrestrial planets relate to their outgassing histories is still in its early stages. Surface conditions and atmospheric composition are controlled by outgassing (together with volatile sinks like atmospheric escape and weathering), while volcanism and tectonics share a profound connection. The case of Mars illustrates that stagnant lid planets could outgas sufficient volatiles to affect the evolution of their surface conditions [127–132]. However, it is often suggested that, on average, volcanism on a planet with plate tectonics would lead to more pronounced outgassing than on a stagnant lid (albeit not necessarily more melt production, see [103]). On the other hand, an episodic lid regime [89] could produce a step-like evolution of surface conditions, with changes occurring in short, sudden bursts, which may be detrimental to life. At an extreme level, some volcanic events, such as Large Igneous Provinces have been suggested to alter surface conditions so strongly that they could cause the end of planetary habitability (see [133] for Venus).

The composition and release of gases during partial melting and volcanic activity depend on the mantle's composition and redox state (see the "Volatile budgets" section). However, uncertainties exist regarding these factors, even for Earth [72,134–139]. The redox state influences the speciation of volatile elements in magmas and, given that these species have different solubilities, the degree to which they are released from magmas (cf. 2, 5A).

The redox state of the mantle (fO_2) is likely inherited from the composition of planetary building blocks and the conditions under which core-mantle equilibrium took place (e.g. [140,141]). However, these initial conditions may be affected by the subsequent tectonic regime experienced by a planet. Notably, it has been proposed that subduction, through the burial of oxidised components such as H_2O, CO_2, and sulfates, has lead to an increase in the oxidation state of Earth's mantle over time, although this remains a topic of debate (e.g., [139,142,143]), while other processes, such as disproportionation of Fe^{2+} into Fe^0 and Fe^{3+}, are also plausible [140,144].

The pressure at which degassing occurs e plays a significant role in determining the composition and mass of the released species [145]. In the case of Venus, with a surface pressure ~90 times greater than Earth's, volcanoes are expected to emit carbon-dominated gases [22], whereas on Earth, volcanoes release more water-rich gases. For large exoplanets with extremely high atmospheric pressures, fewer volatiles would be released [145,146].

The size of a planet also appears as a major factor influencing volatile history. For instance, escape processes depend in a large part on the gravitational pull of the planet, with small planets experiencing stronger loss over time than larger ones (for example, [119]). One must also consider that different species escape at different rates. Lighter species tend to escape more readily, for instance, hydrogen. Heavier species tend to be depleted to a smaller extent. However, energy input from the Sun, atmosphere structure and composition, and photochemical reactions involving each species are critical to quantifying accurately the escape rates of each atmospheric component [147,148]. This affects volatile retention and the whole volatile inventory from accretion to the present day. The specifics of the escape processes vary from planet to planet, not only with size but also atmosphere composition and structure, as well as the presence or absence of a magnetic field (see the "Magnetic field generation: An illustration of the far-reaching effects of mantle dynamics" section). This goes well beyond the scope of this work, and the interested reader is welcome to peruse further review papers, for instance, [147,148], [22] (for a look at the interactions leading to the general volatile inventory), or [80] (for the primitive evolution).

Focusing on internal processes, larger planets also generate more melt over longer periods. Because the heat of formation scales with volume and the heat loss scales with area, larger planets cool more gradually compared to small planets. The composition of the mantle and crust also affects the thermal evolution. For example, stars with $C/O > 0.65$ (for about 30% of stars in the GALAH database) [148] are expected to be hosting carbon-rich rocky exoplanets [149,150] with a graphite lid/crust or a diamond-silicate mantle. The presence of graphite/diamond would speed up planetary cooling by 10 to 100 times, leading to convection modes that are sluggish or stagnant with a low degree of partial melting [151,152]. This difference arises from the 10 to 100 times higher thermal conductivities of graphite and diamond than that of common silicates. The final regime of planets that have lost most of their heat is an inactive stagnant lid (the Moon and Mercury), although not every stagnant-lid planet is geologically "dead". The type of convection or tectonic regime they undergo is also crucial [108,153]. Recent studies have suggested that for stagnant-lid planets, melt production and volatile outgassing are more favorable for planets within the range of 2 to 4 Earth masses [90,154,155].

Observables

Remaining unknowns regarding the interplay between degassing and tectonic regimes still impede efforts to reliably link specific atmospheres types (compositions, species, and mass) to a given regime. Future work should aim at understanding the types of atmospheres that different tectonic regimes can build and their potential for detection.

Magmatic heat flow has been analyzed for various tectonic regimes [108]. Heat flow is closely linked to erupted materials and, consequently, to outgassing. Providing sufficient heat sources, the highest magmatic heat flows are observed among active stagnant-lid planets with volcanism such as Io, which is the most volcanically active body in our Solar System [156]. Io's volcanism efficiently transports approximately 40 times Earth's heat flux from its interior to its surface [157,158]. On the opposite end of the spectrum, plate tectonics and plutonic-squishy lid planets experience the lowest magmatic heat flows. Despite significant differences and uncertainties regarding intrusion to extrusion ratios, these regimes are expected to have relatively similar magmatic activity and outgassing histories compared to other tectonic regimes. Reliable measurements of the heat transport (through surface heat flow) are a major tool of planetary characterization of internal structure and dynamics [110,159,160] but remain delicate, even on Earth [161].

Surface and interior water inventory

Assessing planetary habitability involves understanding the water inventory and its evolution over time. The planet's water retention depends on its evolution during the magma ocean stage, as some or most of the water is retained. On Earth, a significant portion of the delivered water was retained, becoming a fundamental cornerstone in its evolutionary history, which may not have been the case for Mars or Venus (e.g., [22,162]). Collisions could also have affected the planetary climate [163–165].

Tectonic regimes also govern the availability of liquid water on the surface, as well as the weathering of silicate rocks and the burial of marine carbonate rocks into the mantle. These latter mechanisms contribute to climate cooling by removing CO_2 from the atmosphere and the planet's surface, respectively (e.g., [166,167]). In turn, the recycling flux can deliver back volatiles to the mantle that can later be incorporated again in the magma. This mechanism could help prevent depletion of the mantle and ensure continuous outgassing over geological times.

The presence of water has far-reaching effects on mantle rheology, reducing its effective viscosity by several orders of magnitude [168,169]. It also influences melting processes by lowering the mantle solidus temperature (e.g., [170]). Surface water plays a crucial role as well:

1. Hydrated crust facilitates the lubrication of subduction slabs, enabling continuous one-sided oceanic subduction [171,172], a key factor and primary driver of plate tectonics [173].
2. On Earth, after glaciations, the delivery of sediments into oceanic trenches is enhanced, further lubricating subduction and accelerating plate tectonics [174].
3. As a condensable greenhouse gas, water significantly affects surface conditions, up to to runaway greenhouse (refer to the "The Cessation of Habitability" section) [175].
4. The extent of Earth's surface covered by water significantly influences the planet's climate [176–182], which can be linked to the thermoconvective evolution of Earth [183].

Therefore, changes in surface conditions affect the planet's interior. For instance, water (and other gases) influences surface temperature, which, in turn, affects the convection patterns [88,89]. The presence of life and interaction with water further complicate the situation. For instance, it has been suggested that life may trigger plate tectonics [184]. These complex interactions across multiple levels pose challenges for modeling realistic planetary scenarios.

Future research endeavors should therefore also strive to integrate and comprehend the interconnections and feedback loops among tectonics, water circulation, landscapes, climate, and life within diverse tectonic regimes. By doing so, a more complete and detailed depiction of planetary habitability can emerge.

Magnetic field generation: An illustration of the far-reaching effects of mantle dynamics

The generation of a magnetic field by a terrestrial planet is a good example of the links between the interior of a planet and its surface conditions, as well as an argument in favor of considering the planet as a complete system. First, magnetic fields are generated in the planetary core by a magnetic dynamo. Second, enough heat needs to be extracted from the core, through the mantle and out of the solid planet, in order for a dynamo to operate, again demonstrating the importance of considering a planet as a system. Third, the presence of a magnetic field affects the interaction between solar radiation and the planet's atmosphere. In fact, there have been suggestions that a magnetic field may be necessary for a planet to be habitable or, at least, for life to evolve toward a more complex state [1].

The significance of the magnetic field for supporting life is underscored by several arguments. First, it provides protection to a planet's surface by shielding it from harmful solar radiation. In addition, the magnetic field acts as a barrier against some atmospheric (especially water) loss [185]. Among the planets in our Solar System, Earth is the sole planet with a self-generated magnetic field. Mars, on the other hand, bears traces of an extinct magnetic field, while in its ancient crust, which would have been active during the Noachian era, roughly 4 to 3.8 Gyr ago, a period when the planet may have been less arid than it is today. Venus, in contrast, exhibits no evidence of an intrinsic magnetic field in recent history, nor are there any observations indicating remnants of one, which is surprising, given the usual assumption of similar structure and composition compared to Earth [186], and may indicate a radically different thermal evolution.

On Earth (and probably many rocky planets), the iron core is initially fully molten [187], and its cooling is governed by the heat extraction by mantle dynamics. Producing a dynamo requires planetary rotation and convection in a conductive layer. Convection can occur in the liquid part of the core, triggered by the core/mantle temperature contrast or chemical differentiation. As cooling proceeds, solid iron sinks to form a solid inner core while lighter elements (such as H, C, O, Si, and S, for example) rise. The motion of electrically conducting iron in an existing magnetic field induces currents generating a new magnetic field.

In both cases, magnetic fields are an expression of the thermal history and internal structure/composition of terrestrial planets. The rate of heat extraction is required to be large enough for convection to take place. On Earth, it is believed that a minimum heat flow out of the core of 6.9 TW [188] is required, which is nearly within error of Earth's estimated core-mantle boundary (CMB) heat flow of 7 to 17 TW [189]. If Venus had an Earth-like core, the critical heat flow value would probably be around 5 TW [190]. However, different core properties are poorly constrained, crucially the thermal conductivity [186].

In general, stagnant lid regimes are less effective at extracting heat from a planet's interior compared to mobile lid regimes. Consequently, a mobile lid regime, such as plate tectonics, is more likely to sustain long-lasting convection in the core. This disparity may be exemplified by the contrasting situations of Earth and Mars [110], where Mars has not generated a magnetic field for approximately 4 Ga, due to its still liquid core [191] and the lack of efficient heat extraction after the primitive period. On Venus, other reasons have been proposed to be behind the absence of a magnetic field. These include an already fully solidified core [192] or a stably stratified core [186,193] resistant to convective motion (possibly due to the absence of remixing caused by a giant impact, such as Earth is thought to have had leading to the Moon formation).

However, recent observation of the atmospheric escape rates of O^+ ion on Mars, Venus, and Earth suggests that they are rather similar. Loss rates ranges from 2×10^{24} to 4×10^{24} O^+ s^{-1} for Mars, to 1×10^{24} to 4×10^{24} O^+ s^{-1} for Venus, and 5×10^{24} to 5×10^{25} O^+ s^{-1} for Earth (see [194] and references therein). This implies that perhaps the magnetic field is less important than previously thought regarding atmospheric escape.

It has been suggested that a magnetic field can extend much further away from the surface of the planet than the atmosphere itself, which causes a magnetized planet to intercept a larger portion of stellar energy when compared to an unmagnetized one [195]. That additional energy would power inflated escape rates, possibly higher than for unmagnetized planets (e.g., [196]). This postulate has been challenged by Tarduno et al. [185], who point out poorly constrained return fluxes and the variability of escape rates with the solar flux and geodynamo even for a single planet [194].

An undeniable effect of self-generated magnetic fields is that they modify the preferential escape location: Ions flow along magnetic field lines under the effect of the solar extreme ultraviolet (EUV) flux. With an induced magnetic field, ions could be lost in the magnetotail, as magnetic lines are open. With self-generated magnetic field, magnetic lines in the tail reconnect to the planet, preventing ions to escape. However, the cusps of the magnetic field contain open-field lines and allow solar radiation to excite ions deep in the atmosphere [197] and accelerate them downtail through the lobes of the field, leading to their escape. This was further illustrated by models of magnetized planets with a high obliquity.

It is possible that what differs between magnetized and unmagnetized planets is the response of magnetospheres/ionospheres to intense events, like coronal mass ejections that are far more numerous during planetary early evolution [198]. To summarize, current research casts doubt onto the classical view of the magnetic field as a shield against escape but highlights the complexity of its interactions with solar radiation. The net result for terrestrial planets remains uncertain.

The Cessation of Habitability

Liquid water and carbon cycling

The breakdown in one of the many processes that enable long-term habitable conditions on Earth can drive the cessation of habitability. Liquid water, an energy source, CHONPS, transition metals, and relevant geochemical conditions enable life to originate and evolve. The sustenance of a core-dynamo-driven magnetic field protects Earth's biosphere from cosmic rays and charged solar particles. The steady interior evolution and plate tectonics drive outgassing and recycling of volatiles on Earth. The presence of oceans, the hydrologic cycle, and the carbonate-silicate cycle (inorganic carbon cycle) allow for temperate climates by regulating CO_2, a vital greenhouse gas. Among these factors, the presence of liquid water on the surface and the geochemical cycling of CO_2 are key in ensuring habitable conditions on billion-year time scales.

Unlike Venus, where most of the near-surface CO_2 is present in the atmosphere, on present-day Earth, more than 99% of near-surface CO_2 is expected to be locked up in the form of carbonates, courtesy of the carbonate-silicate cycle (e.g., [199–203]). A number of steps are involved in recycling CO_2 between the atmosphere and the interior of Earth [204–206]. The continental weathering of silicate rocks [207,208] and subsequent seafloor precipitation of carbonates locks up CO_2 in solids [209]. Eventual subduction of carbonates removes CO_2 from the atmosphere–ocean system [205]. In the upper mantle, due to metamorphism, carbonates break down into silicates and release CO_2 back into the atmosphere [210].

To sustain CO_2 cycling, several factors are expected to be crucial, including the hydrologic cycle, fresh silicate rocks, and a planetary orbit within the HZ [211]. The incident stellar radiation (instellation) on Earth has gradually increased over the past 4.5 Gyr because of the Sun's evolution on the main sequence of the Hertzprung–Russell diagram. The faint young Sun paradox suggests that early Earth must have been frozen to maintain an energy balance between solar irradiation and outgoing longwave radiation for early Earth with similar albedo and atmospheric greenhouse effects as modern Earth [212]. The contrary evidence of an active hydrologic cycle during the Archean calls for a mechanism to ensure temperate climates during Earth's history [213]. The carbonate-silicate cycle-driven steady decline in CO_2 partial pressure (P_{CO_2}) from approximately 0.1 bar (uncertainty range, 0.01 to 1 bar) during the Archean to a preindustrial value of 0.28 mbar has been suggested to be a solution for the faint young Sun paradox [207,214]. During the Phanerozoic, the biosphere has further enhanced the efficiency of CO_2 cycling by adding an organic component to the carbonate-silicate cycle [210,211]. Which factors can break down the carbonate-silicate cycle? This question is key because it invokes a number of processes of habitability cessation that are directly or indirectly involved in the carbonate-silicate cycle.

Silicate weathering and carbonate precipitation

If silicate weathering were not operational on Earth, the partial pressure of atmospheric CO_2 accumulated because of volcanism would already become 1 bar in 20 million years (Myr) [207]. This would rapidly increase the surface temperature due to the greenhouse effect of CO_2. Fortunately, that is not the case on Earth. The balance between weathering and outgassing on a shorter time scale of 0.1 to 1 Myr ensures a quasi-steady state in atmospheric CO_2 [215]. On a longer time scale of 0.1 to 1 Gyr, all planetary CO_2 reservoirs can attain equilibrium and prevent a runaway greenhouse state [8,205]. The negative feedback of silicate weathering ensures that the accumulation of CO_2 due to outgassing will be balanced by the loss of CO_2 due to silicate weathering, thereby maintaining moderate amounts of steady-state CO_2 on shorter time scales. On longer time scales, the intensity of silicate weathering must gradually increase with stellar luminosity.

The intensity of weathering is limited by the availability of weathering reactants, i.e., water, silicate rocks, and CO_2 [207]. This leads to 3 limits of weathering [166,208,216]. If fresh silicate rocks cannot be supplied by plate tectonics, weathering becomes rock-supply-limited W_{su} [8,166]. However, plate tectonics may not be necessary for carbonate burial and recycling (e.g., CO_2 cycling for stagnant-lid planets [217,218]). From the perspective of the limited transport of hydrologically important flowpaths to fresh minerals leading to dilute water streams, supply-limited weathering is also called transport-limited [208,219]. Weathering becomes kinetically limited W_{ki} when the time available for weathering reactions is shorter than the reaction equilibrium time scale [203,204]. If there is limited availability of water (e.g., arid climates), weathering becomes thermodynamically limited W_{th} [216,220,221]. Contrary to the kinetic limit, at the thermodynamic limit, weathering decreases as a function of temperature, resulting in a positive weathering feedback, whose impact on climate stability is not understood yet [221]. To accurately model CO_2 cycling of temperate planets beyond modern-Earth conditions, Hakim et al. [221] and Graham and Pierrehumbert [222] demonstrate the need to use a multiregime weathering equation ([216], adapted formulation):

$$W_{total} = \left(\frac{1}{W_{th}} + \frac{1}{\left(\frac{1}{W_{ki}} + \frac{1}{W_{su}} \right)^{-1}} \right)^{-1}. \tag{7}$$

With such a formulation, the kinetic weathering regime [207] is recovered when $W_{ki} < W_{su}$ and $W_{ki} < W_{th}$, which is largely the case for modern Earth. The weathering is rock-supply-limited when $W_{su} < W_{ki}$ and $W_{su} < W_{th}$, and the weathering is at the thermodynamic limit when $W_{th} < (1/W_{ki} + 1/W_{su})^{-1}$.

The type of silicate rocks also affects the intensity of weathering. For example, ultramafic and mafic rocks contribute up to 10 times more to weathering than felsic rocks [221,223]. Rocky exoplanets are expected to show a large diversity in silicate rock types, as studies using stellar elemental abundances suggest [224,225]. Other planets may feature a wide range of gas/surface chemical reactions, which may or may not be conducive to habitability. For example, it is thought that oxidation reactions on Venus could consume or buffer oxidized components of the atmosphere, such as H_2O, SO_2, or CO_2 [22,226].

As the products of silicate weathering reactions are transferred to oceans by rivers, carbonates precipitate and settle on the seafloor for eventual subduction into the mantle. Carbonate precipitation is the following step after silicate weathering. On present-day Earth, the precipitation of calcium carbonates (calcite and aragonite) occurs almost instantaneously compared to silicate weathering [205]. The precipitation of carbonates is generally neglected in carbon cycle box models because it is not the rate-limiting state among the two. However, even when continental silicate weathering takes place but the oceans are too deep, carbonate precipitation cannot occur. Below a certain ocean depth, known as the carbonate compensation depth (CCD), carbonates are unstable because of high pressures and dissolve back into the ocean [209,227]. To ensure the availability of carbonates for subduction, in principle, the CCD should be deeper than the ocean depth, which is the case for modern Earth. The CCDs for the carbonates of calcium, magnesium, and iron become deeper at a higher P_{CO2} and a higher temperature, thereby favoring CO_2 cycling for diverse carbonates [228].

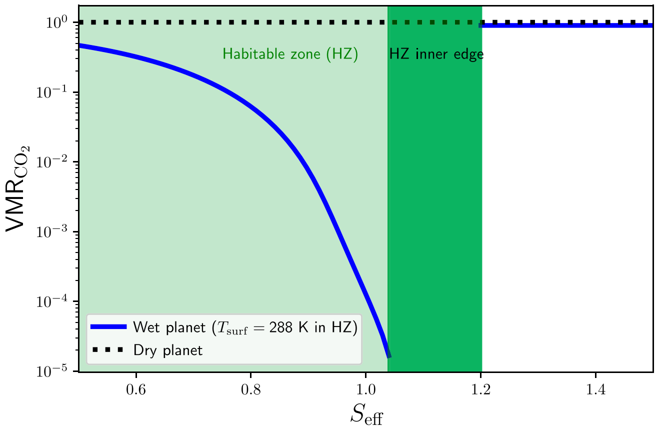

Fig. 5. $\mathrm{VMR}_{\mathrm{CO2}}$ of 2 exoplanet types as a function of instellation normalized to that of modern Earth (S_{eff} =1). A wet planet, i.e., an Earth-like planet with surface water oceans, carbonate-silicate cycling, and an N_2-dominated atmosphere, can potentially maintain a constant surface temperature T_{surf} = 288 K by decreasing $\mathrm{VMR}_{\mathrm{CO2}}$ in response to an increase in S_{eff} (calculated using the climate model of [307]). The inner HZ edge is shown as an extended zone because 1- and 3-dimensional climate models predict the inner edge to be between S_{eff} = 1.05 to 1.2 [15,231–234]. At higher instellations beyond the inner HZ edge, water evaporation/loss followed by near-surface CO_2 desiccation will make the atmosphere of the wet planet CO_2-dominated exhibit a jump in $\mathrm{VMR}_{\mathrm{CO2}}$ [222,239,308]. For a CO_2-dominated dry planet, the inner edge of HZ is not expected to affect $\mathrm{VMR}_{\mathrm{CO2}}$.

In the absence of continents, seafloor weathering and CO_2 dissolution on ocean worlds can also stabilize the long-term climate [219,229,230]. The total liquid water inventory and topography of terrestrial planets (which depend on outgassing and mantle dynamics) can strongly affect the drawdown of CO_2.

Observational tests

Inside the HZ for a wet planet, the carbonate-silicate cycle can be assumed to result in a constant surface temperature (e.g., T_{surf} = 288 K; Fig. 5). Such a stable temperature can be sustained by ensuring a higher P_{CO2} at lower instellation due to a lower intensity of silicate weathering and a lower P_{CO2} at higher instellation due to a higher intensity of silicate weathering. At sufficiently high instellations (higher than 1.05 to 1.2 S_{eff}, where S_{eff} = 1 denotes modern Earth insolation) [15,231–234], the weathering feedback cannot mitigate climate warming (above a fixed energy deposition limit). This leads to the evaporation of water and an increase in the atmospheric water vapor abundance beyond the regular water vapor positive feedback [211]. The intricate balance between climate feedbacks and geochemical cycling is key to avoiding such an irreversible increase in surface temperature by runaway greenhouse due to the water vapor feedback mechanism or by an additional greenhouse gas that pushes the surface temperature beyond the critical point of water on its own (the Simpson–Nakajima and Komabayasi–Ingersoll limits, respectively) [235–238]. The orbital distance at which the runaway greenhouse occurs is thought to define the inner HZ edge [15,16]. A bimodal distribution in the CO_2 volume mixing ratio ($\mathrm{VMR}_{\mathrm{CO2}}$) across the inner HZ edge for wet rocky planets is statistically observable [222,239,240]. Bean et al. [239] suggested that 20 planets could be sufficient to identify the $\mathrm{VMR}_{\mathrm{CO2}}$ jump at the inner HZ edge (Fig. 5). On the other hand, Lehmer et al. [241] made statistical predictions for the planets inside the HZ. If the probability of observed planets matches a log-uniform distribution of P_{CO2} regulated by the carbonate-silicate cycle, future telescopes need to observe at least 83 HZ planets [241]. Schlecker et al. [242] suggest that a high-precision transit photometry survey, such as European Space Agency (ESA) PLATO [243], can observationally identify the inner HZ edge $\mathrm{VMR}_{\mathrm{CO2}}$ jump for a sample of 100 exoplanets with at least 10% exhibiting runaway climates.

It is important to note that an alternative climate state with CO_2 as clathrate hydrate or liquid CO_2 can also be stable for billions of years [244]. The critical absorbed flux can be crossed because of an increase in stellar radiation or changes in albedo, for instance. The moist atmosphere has a limit to how much energy it can radiate into space, which will further increase the surface temperature until the atmosphere radiates in the near-infrared. It is possible that a runaway greenhouse could lead to partial melt of the surface for large enough surface liquid water inventory (about 1,400 K for a few tens of bar) [245]. This positive feedback can persist with almost no bounds leading to a runaway greenhouse. Venus is suspected to have undergone such a runaway greenhouse during its past [175,200]. Even on Earth, such a runaway greenhouse is expected to occur in the future as the Sun becomes brighter on its main sequence evolution.

Current and Future Missions

Solar System missions

To this date, no space mission has ever found definitive evidence for life in the Solar System, beyond Earth, but habitable environments have been suggested for specific locations and times. Ancient Mars, ancient Venus, and outer Solar System icy satellites are all interesting targets. Direct observation of habitable conditions is difficult, and space missions tend to investigate concomitant properties of the planet and environment to help understand planetary evolution.

Mars

- Mars is the most explored planet in the Solar System, apart from Earth. After negative results during the Viking mission [246], it was recognized that present-day Mars is not favorable to life, despite minute signs that habitability might not be far off martian conditions. Any habitable phase would have occurred more than 3.8 Gyr ago.

Further exploration operated under the “follow the water” guideline and attempted to characterize the ancient geology of Mars and put time constraints on water-related, geological, and mineralogical features (e.g., [247–249]). Mars, once, had more water (liquid, at times) than at present, even if it remains uncertain whether habitable conditions were stable or episodic or if the climate was warm or cold (see [250,251], for instance). Further investigation has targeted interactions between the interior and the atmosphere. Phoenix (which arrived on Mars in 2008) has investigated polar region rhegolith and found that it has interacted little with liquid water in recent history [252]. The Curiosity rover (landed in 2012) and Perseverance (landed in 2018) looked into the possible subsurface origin of methane [253,254] and provided data regarding surface mineralogy, past traces of water, the lower atmosphere, and geology. Finally, Tianwen-1 (China National Space Administration) landed in 2021 and notably explored the relation between surface features, the past presence of water and the nature of the martian surface. Insight (2018–2022) aimed to cover the deep interior of Mars and its structure, composition and dynamics, in particular, using a seismometer [191,255]. The mission recorded hundreds of Mars-quakes and constrained the size of the core and its fraction of light elements. Future Mars-related astrobiology missions include multiple plans for a Mars Sample Return (NASA), as many questions require direct analyses of martian material [256,257]. Plans for the Tianwen-3 mission (China National Space Administration) also include sample return (launch scheduled for 2028). Finally, the 2-part ExoMars mission (2016 and 2028 launches) attempts to investigate biosignatures, water distribution, carbon isotopes and their possible biologic origins.

Venus

- Venus has been relatively little explored compared to other targets in the Solar System.

Early remote observation, Mariner 2 and the soviet Vega and Venera landers (in the 60s and the 70s) revealed the crushing 92-bar surface pressure and ≈740-K surface temperature caused by its massive CO_2 atmosphere (see [258], for a detailed overview). Ancient habitable Venus has nonetheless been suggested [259] but is debated (see [22,24]) and depends on very specific atmospheric conditions [260]. Present-day habitability of the atmosphere near the cloud layer has been suggested [261] but remains unconfirmed, while recent claims of the observation of phosphine [262] in the atmosphere have again triggered heated debate [263–267]. Beyond invaluable but imprecise in situ measurements (Vega and Venera), Magellan (1990) and Venus Express (2006) have been the source of most of the current scientific knowledge of the surface and atmosphere of Venus [258]. Additional observation was performed by Akatsuki (2015) and the Parker Solar Probe (2018), as well as other missions en route toward other targets [268].There is a general consensus that we still know very little about Venus' interior structure and composition, and lower atmosphere. However, the evolution of Venus' interior and how it may have caused a divergent evolution compared to Earth are critical research avenues for habitability. This caused a resurgence of interest in Venus with several high-profile missions planned in the next decades: DaVinci and VERITAS (USA), EnVision (EU), VENERA-D (RU), Shukrayaan-1 (India), VOICE (China). All 6 missions are designed to investigate the past of the planet and the processes that affect the evolution of its surface conditions and volcanic processes. Ancient traces of water are also investigated, throughout the atmosphere, and near possibly ancient material (the tesserae) [159,269,270]. The main goal of those missions is not just to know whether Venus was habitable but to understand why it is not at present day, so that this knowledge can be applied more broadly [22,271].

The icy moons of Jupiter and Saturn

- The icy moons of Jupiter and Saturn are considered important targets for astrobiology missions. The Cassini Huygens mission opened the door to further exploration of the satellites of giant outer Solar System planets. The landing on Titan and discovery of its surface and liquid hydrocarbon lakes has seen suggestion of methanogenic life [272]. Io, Enceladus, Ganymede, and Europa are all suspected of harboring (possibly habitable) interior liquid water oceans beneath their icy surface [273–276]. Future missions include ESA's Juice toward Europa, Ganymede, and Calisto [277]. Europa is also the target of NASA's Europa Clipper [278]. NASA's second mission, Dragonfly [279,280], would place a rotorcraft on Titan to study biosignatures and chemistry.

Exoplanet missions

In the context of detecting and, even more importantly, characterizing the (atmospheric) properties of temperate, terrestrial exoplanets, it is important to consider a few aspects:

- M-type stars versus solar-type stars: Depending on the effective temperature and luminosity of the host star (note that we are primarily concerned with main-sequence stars here), the orbital period (and hence mean orbital separation) of HZ planets varies significantly, ranging from up to a few hundred days for stars similar or slightly hotter than the Sun, to less than a few tens of days for cool, low-luminosity M-type dwarf stars. This has important implications, which techniques can be used for atmospheric investigations of the objects.
- Transit spectroscopy and secondary eclipse observations versus direct detection methods: Currently, the most successful approach to observing the atmospheres of exoplanets are transit spectroscopy and secondary eclipse measurements (for an introduction, see [281]). These techniques have a strong bias toward large, close-in planets. In addition, while the James Webb Space Telescope (JWST) and ESA's Ariel mission [282] will revolutionize our understanding of exoplanet atmospheres for hot/warm gas dominated planets, they will not be able to probe a large sample of rocky, temperate exoplanets. In the most favorable cases, which are small planets transiting very low-mass stars (such as the well-known Trappist-1 system) [283], JWST may be able to tell whether these objects have an atmosphere at all, which will be an important milestone in exoplanet science (for first results, see [284,285]), but an in-depth characterization and comprehensive inventory of atmospheric constituents will likely be out of scope (e.g., [206,286]). The reason is very simple: Despite the large 6.5-m primary mirror, the favorable planet-star brightness ratio, the proximity of the Trappist-1 system of only ≈12 pc, and the ability to observe and stack several transits due to the very short orbital periods of the planets, the signal-to-noise ratio of the data will remain limited. This is the reason why, in the future, a main focus will be on direct detection methods that are largely independent from orbital properties and do not require an exoplanet to transit its host star (for completeness we mention here, the ambitious and inspiring Nautilus project that seeks to search for biosignatures in the atmospheres of hundreds of transiting terrestrial exoplanets with a fleet of low-cost spacecraft [281]); the vast majority of exoplanets does, in fact, not transit in front of their stars!
- Reflected light versus thermal emission: Accepting that direct detection techniques are a promising way to probe many of the temperate, terrestrial exoplanets within 10 to 25 pc from the Sun, one needs to realize that there are 2 ways to directly detect light from these objects: One can either probe the starlight reflected by the planets (this is typically done at ultraviolet, optical, and near-infrared wavelengths between ~0.2 and ~2.5 μm), or one can

probe the planets' intrinsic thermal emission (which, for temperate objects similar to the terrestrial planets in the Solar System, can be done at mid-infrared wavelengths between ~3 and ~20 μm).

With currently available telescopes (on ground and in space), we cannot directly detect small, HZ exoplanets. It is technically not feasible as none of the existing facilities or instruments provide the full combination of required spatial resolution, contrast performance, and sensitivity. For the required contrast, we remind the reader that in reflected light at optical wavelengths Earth is a factor of $\approx 10^{10}$ fainter than the Sun and at a wavelengths of ~11 μm, where the thermal emission of Earth peaks, the flux difference still amounts to $\approx 10^{7}$ (e.g., [287]). For planets orbiting low-luminosity M-dwarfs, these values are typically a factor of 100 lower.

Considering the points listed above, it becomes clear that a number of new instruments and missions will be needed for a comprehensive investigation of a large number of temperate terrestrial exoplanets orbiting various kinds of host stars and across a large wavelength range:

- Probing nearby exoplanets orbiting low-luminosity M-dwarfs in reflected light will be the primary discovery space of optimized instruments installed at extremely large telescopes (ELTs) (e.g., [288]). With their 30- to 40-m primary mirrors, they will provide sufficient spatial resolution to spatially separate the signal from the planet from that of its host star on the detector, and the combination of extreme adaptive optics systems with high-resolution spectrographs will deliver the required contrast performance [289].
- Probing nearby exoplanets orbiting Sun-like stars in reflected light will require dedicated space missions. The stringent contrast requirement (see above) can likely not be achieved by ground-based instruments due to the disturbing effects from Earth atmosphere. In fact, a large, single aperture exoplanet imaging mission was recently recommended in the context of NASA's 2020 Astrophysics Decadal Survey (https://www.nationalacademies.org/our-work/decadal-survey-on-astronomy-and-astrophysics-2020-astro2020). The main science driver for this so-called Habitable World Observatory (HWO) is to detect and investigate ≈25 Earth-like planets around nearby Sun-like stars in reflected light. While the exact wavelength coverage is still under investigation—in particular, the short- and long-wavelength cutoff in the ultraviolet and near-infrared will be carefully examined—the aperture size shall be ≈6 m and hence in between those of the Large UV/Optical/IR Surveyor (LUVOIR) and Habitable Exoplanet Observatory (HabEx) mission concepts, which have been studied in preparation for the Decadal Survey [290,291]. We remind the reader that the Hubble Space Telescope has a primary mirror of 2.4 m. The oxygen A-band (759 to 770 nm) is the main diagnostic for HWO to search for indications of biological activity in exoplanet atmospheres, but absorption bands of water, and potentially carbon dioxide and methane, shall also be accessible. First studies quantifying how well Earth-like planets can be characterized by a future reflected light mission like HWO have been carried out [292].
- Finally, probing the thermal emission of nearby exoplanets orbiting any type of host star, requires a large, space-based nulling interferometer. Even with the 39-m ELT of the European Southern Observatory, the thermal background noise at mid-infrared wavelength for ground-based observations will be prohibitively high so that only very few of the nearest stars can be imaged directly in the search of temperate terrestrial exoplanets [293,294]. From space, however, a mission like the Large Interferometer For Exoplanets [295] could detect hundreds of nearby exoplanets including dozens similar to Venus and Earth in terms of size and energy influx [296,297]. An exoplanet's thermal emission allows for a robust characterization of its atmospheric properties such as pressure–temperature profile and composition [298] as it is less affected by disturbing effects such as clouds [299] compared to observations in reflected light. In the case of a Large Interferometer For Exoplanets-like mission, habitable conditions on an Earth-like planet can readily be inferred from moderate resolution mid-infrared spectra throughout most of Earth's history [300,301], and the mid-infrared wavelength range provides access to a variety of atmospheric biosignatures gases (e.g., [302]). We emphasize that the direct detection of temperate terrestrial exoplanets at mid-infrared wavelengths with the goal to identify habitable—and potentially even inhabited—exoplanets was given very high scientific priority in the Voyage 2050 program of the ESA and has been identified as a potential topic for a future L-class mission in ESA's science program (https://www.cosmos.esa.int/web/voyage-2050).

Looking at the list above, one can be hopeful that in the coming 20 to 30 years, we will indeed have a suite of instruments and missions that strongly complement each other in the search for habitable worlds beyond the Solar System. These projects are not in competition with each other; in addition to the unique scientific diagnostic that each of them possesses, the combined power of reflected light and thermal emission data will provide a truly holistic picture of temperate terrestrial exoplanets orbiting a wide range of stellar types in our cosmic neighborhood.

Conclusions and Summary

Habitability is a complex notion dependent on many interwoven processes. The qualitative definition of habitability being a set of conditions that are conducive to life is generally insufficient for any precise assessment or prediction. On the other hand, more quantitative definitions are strongly biased by the only example of a unquestionably habitable world we possess.

At its core, the habitability of a planet is generally restricted to assessing the suitability of the conditions at its surface for self-sustaining biotic reactions to occur; that is, the response of a planetary atmosphere to the energy input (such as stellar radiation, for example). However, that response depends on the planet's time-integrated atmospheric structure and composition, which are governed by (a) the initial availability of life-essential elements in the planetary body and (b) a myriad of interacting mechanisms and feedback processes. The later can operate both inside the atmosphere (chemistry and dynamics) or outside (volatile or energy exchange with the planetary interior). The fate of water is especially important for life (as we know or imagine it, at least) and constitutes the first-order guide for habitability models and detection.

The relative importance of these different processes changes depending on the stage of planetary evolution and time. Two of the main drivers of changes are the evolution of the host star and that of the silicate planet. In this review, we have discussed some of the most high-profile questions regarding the role of the interior of terrestrial planets in controlling habitability, such as the conditions for its onset, the role of interior dynamics, and how it is maintained and could cease.

The nature of terrestrial planets, by definition, is distinct from those of gas giants that have nearly their full complement of stellar gases. By contrast, terrestrial planets are relatively "dry" [47]. Therefore, the masses of atmospheres around terrestrial planets are products of the bulk composition of the planet and the prevailing pressure–temperature conditions under which the atmosphere formed and evolved through time. Given that >99.95% of the bulk silicate mass of planets resides in rocks [38], the proportions of O to other rock-forming elements, namely, Fe, Mg, and Si, control the nature of atmospheres around rocky planets through the fO_2. As such, a planet's interior governs the long-term evolution of the atmosphere by exchanging mass (in the form of volatile species) and heat with it. The net flux of outgassing (e.g., volcanic exhalation) and ingassing (burial of volatile elements into the mantle and crust) defines the bulk composition of the atmosphere (and surface) and thus its water inventory (see [35,40]). We have underlined the importance of the magma ocean stage in setting the initial conditions that can lead to habitable surface conditions (see the "The Onset of Habitability" section). Later, habitability is maintained or fails because of a delicate balance in volatile exchange. On Earth, the relevant species are water and CO_2 owing to the relatively oxidised nature of its mantle (ΔIW+3.5 at present in the upper mantle) [144], but on other planets, the list could contain other gaseous species, namely, H_2 or CO on more reduced planets and SO_2 on more oxidizing planets.

This field of investigation is still in its infancy, as results of the past few decades have highlighted the vast complexity of the interactions leading to volatile exchange (e.g., [40,48,76,83,303]). They involve all layers of the planet, from the core to the upper atmosphere, in addition to the more obvious atmosphere/mantle interface. This is an inherently multidisciplinary endeavour that requires a deep understanding of all the parts of the volatile cycle. In short, understanding habitability requires us to understand planetary evolution as a whole, an equally exciting and daunting prospect.

Current scientific efforts include many missions especially focused on the theme of habitability, either in our Solar System or toward exoplanets. It is tempting to look for the safest targets, selecting objects similar to Earth, but it is better to cast a wide net and see what comes up. Many missions attempt this with planets that have very different conditions from Earth, such as Venus, for example, to understand the mechanisms of planetary evolution or for exoplanets to build-up statistics beyond our small Solar System. After all, there is no guarantee that Earth is the unique blueprint for habitability. At present, we still understand the topic far too little to make solid predictions. Instead, we would reap tremendous benefits form observation of the diversity of possible planetary evolution outcomes.

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References

1. Cockell CS, Bush T, Bryce C, Direito S, Fox-Powell M, Harrison JP, Lammer H, Landenmark H, Martin-Torres J, Nicholson N, et al. Habitability: A review. *Astrobiology*. 2016;16(1):89–117.
2. Hinkel NR, Timmes F, Young PA, Pagano MD, Turnbull MC. Stellar abundances in the solar neighborhood: The hypatia catalog. *Astron J*. 2014;148(3):Article 55.
3. Seckbach J. *Genesis-in the beginning: precursors of life, chemical models and early biological evolution*. Berlin (Germany): Springer Science & Business Media; 2012.
4. Stüeken EE, Anderson RE, Bowman JS, Brazelton WJ, Colangelo-Lillis J, Goldman AD, Som SM, Baross JA. Did life originate from a global chemical reactor? *Geobiology*. 2013;11(2):101–126.
5. Seager S. Exoplanet habitability. *Science*. 2013;340(6132):577–581.
6. Laskar J, Joutel F, Robutel P. Stabilization of the Earth's obliquity by the Moon. *Nature*. 1993;361(6413):615–617.
7. Jellinek AM, Jackson MG. Connections between the bulk composition, geodynamics and habitability of Earth. *Nat Geosci*. 2015;8:587–593.
8. Foley BJ. The role of plate tectonic–climate coupling and exposed land area in the development of habitable climates on rocky planets. *Astrophys J*. 2015;812(1):36.
9. Brownlee D, Ward P. *Rare Earth*. NY: Copernicus; 2000.
10. Lenardic A, Seales J. Different is more: The value of finding an inhabited planet that is far from Earth2.0. *Astrobiology*. 2019;19(11):1398–1409.
11. Dartnell L. Biological constraints on habitability. *Astron Geophys*. 2011;52(1):1.25–1.28.
12. Lammer H, Bredehöft J, Coustenis A, Khodachenko ML, Kaltenegger L, Grasset O, Prieur D, Raulin F, Ehrenfreund P, Yamauchi M, et al. What makes a planet habitable? *Astron Astrophys Rev*. 2009;17(2):181–249.
13. Lingam M. A brief history of the term 'habitable zone' in the 19th century. *Int J Astrobiol*. 2021;20(5):332–336.
14. Zuluaga JI, Salazar JF, Cuartas-Restrepo P, Poveda G. The habitable zone of inhabited planets. *Biogeosci Discuss*. 2014;11(6):8443–8483.
15. Kopparapu RK, Ramirez R, Kasting JF, Eymet V, Robinson TD, Mahadevan S, Terrien RC, Domagal-Goldman S, Meadows V,

Deshpande R. Habitable zones around main-sequence stars: New estimates. *Astrophys J*. 2013;765(2):Article 131.
16. Kasting JF, Whitmire DP, Reynolds RT. Habitable zones around main sequence stars. *Icarus*. 1993;101(1):108–128.
17. Pierrehumbert R, Gaidos E. Hydrogen greenhouse planets beyond the habitable zone. *Astrophys J Lett*. 2011;734(1):L13.
18. Rugheimer S, Kaltenegger L, Zsom A, Segura A, Sasselov D. Spectral fingerprints of Earth-like planets around FGK stars. *Astrobiology*. 2013;13(3):251–269.
19. Rugheimer S, Kaltenegger L. Spectra of Earth-like planets through geological evolution around FGKM stars. *Astrophys J*. 2018;854(1):19.
20. Turnbull MC, Tarter JC. Target selection for SETI. I. A catalog of nearby habitable stellar systems. *Astrophys J Suppl Ser*. 2003;145(1):181.
21. Driscoll P, Bercovici D. Divergent evolution of Earth and Venus: Influence of degassing, tectonics, and magnetic fields. *Icarus*. 2013;226(2):1447–1464.
22. Gillmann C, Way MJ, Avice G, Breuer D, Golabek GJ, Höning D, Krissansen-Totton J, Lammer H, O'Rourke JG, Persson M, et al. The long-term evolution of the atmosphere of Venus: Processes and feedback mechanisms. *Space Sci Rev*. 2022;218:56.
23. Checinska Sielaff A, Smith SA. Habitability of Mars: How welcoming are the surface and subsurface to life on the red planet? *Geosciences*. 2019;9(9):361.
24. Westall F, Höning D, Avice G, Gentry D, Gerya T, Gillmann C, Izenberg N, Way MJ, Wilson C. The habitability of Venus. *Space Sci Rev*. 2023;219:17.
25. Albarede F. Volatile accretion history of the terrestrial planets and dynamic implications. *Nature*. 2009;461(7268):1227–1233.
26. Sossi PA, Stotz IL, Jacobson SA, Morbidelli A, O'Neill HSC. Stochastic accretion of the Earth. *Nat Astron*. 2022;6(8):951–960.
27. Wetherill G. Provenance of the terrestrial planets. *Geochim Cosmochim Acta*. 1994;58(20):4513–4520.
28. Jacobson SA, Morbidelli A. Lunar and terrestrial planet formation in the grand tack scenario. *Philos Trans R Soc A Math Phys Eng Sci*. 2014;372(2024):Article 0174.
29. Izidoro A, Dasgupta R, Raymond SN, Deienno R, Bitsch B, Isella A. Planetesimal rings as the cause of the Solar System's planetary architecture. *Nat Astron*. 2022;6:357–366.
30. Morbidelli A, Baillié K, Batygin K, Charnoz S, Guillot T, Rubie DC, Kleine T. Contemporary formation of early Solar System planetesimals at two distinct radial locations. *Nat Astron*. 2022;6:72–79.
31. Johansen A, Lambrechts M. Forming planets via pebble accretion. *Annu Rev Earth Planet Sci*. 2017;45:359–387.
32. Fegley B Jr, Lodders K, Jacobson NS. Volatile element chemistry during accretion of the Earth. *Geochemistry*. 2020;80(1):Article 125594.
33. Nimmo F, Price G, Brodholt J, Gubbins D. The influence of potassium on core and geodynamo evolution. *Geophys J Int*. 2004;156(2):363–376.
34. McCubbin FM, Barnes JJ. Origin and abundances of H_2O in the terrestrial planets, Moon, and asteroids. *Earth Planet Sci Lett*. 2019;526:Article 115771.
35. Marty B. The origins and concentrations of water, carbon, nitrogen and noble gases on Earth. *Earth Planet Sci Lett*. 2012;313–314:56–66.
36. Dauphas N. The isotopic nature of the Earth's accreting material through time. *Nature*. 2017;541(7638):521–524.
37. Broadley MW, Bekaert DV, Piani L, Füri E, Marty B. Origin of life-forming volatile elements in the inner Solar System. *Nature*. 2022;611(7935):245–255.
38. Palme H, O'Neill HS. *Treatise on geochemistry.* Oxford: Elsevier; 2014.
39. Wang Z, Becker H. Ratios of S, Se and Te in the silicate earth require a volatile-rich late veneer. *Nature*. 2013;499(7458):328–331.
40. Hirschmann MM. Comparative deep earth volatile cycles: The case for c recycling from exosphere/mantle fractionation of major (H_2O, C, N) volatiles and from H_2O/Ce, CO_2/Ba, and CO_2/Nb exosphere ratios. *Earth Planet Sci Lett*. 2018;502:262–273.
41. Rubey WW. Geologic history of sea water: An attempt to state the problem. *Geol Soc Am Bull*. 1951;62(9):1111–1148.
42. Li Y, Vočadlo L, Sun T, Brodholt JP. The Earth's core as a reservoir of water. *Nat Geosci*. 2020;13:453–458.
43. Hirose K, Wood B, Vočadlo L. Light elements in the Earth's core. *Nat Rev Earth Environ*. 2021;2:645–658.
44. Speelmanns IM, Schmidt MW, Liebske C. The almost lithophile character of nitrogen during core formation. *Earth Planet Sci Lett*. 2019;510:186–197.
45. Grewal DS. Origin of nitrogen isotopic variations in the rocky bodies of the Solar System. *Astrophys J*. 2022;937(2):123.
46. Blanchard I, Rubie DC, Jennings ES, Franchi IA, Zhao X, Petitgirard S, Miyajima N, Jacobson SA, Morbidelli A. The metal–silicate partitioning of carbon during Earth's accretion and its distribution in the early Solar System. *Earth Planet Sci Lett*. 2022;580:Article 117374.
47. Aston F. The rarity of the inert gases on the Earth. *Nature*. 1924;114(2874):786.
48. J. Fegley, B., L. Schaefer, *The atmosphere - Treatise on geochemistry*. 2nd Ed. J. Farquhar, D. Canfield, J. Kasting, editors (Elsevier, 2014).
49. Halliday AN. The origins of volatiles in the terrestrial planets. *Geochim Cosmochim Acta*. 2013;105:146–171.
50. Olson PL, Sharp ZD. Nebular atmosphere to magma ocean: A model for volatile capture during earth accretion. *Phys Earth Planet Inter*. 2019;294:Article 106294.
51. Péron S, Moreira M, Agranier A. Origin of light noble gases (He, Ne, and Ar) on Earth: A review. *Geochem Geophys Geosyst*. 2018;19(4):979–996.
52. CMO'D Alexander, Bowden R, Fogel ML, Howard KT, CDK H, Nittler LR. The provenances of asteroids, and their contributions to the volatile inventories of the terrestrial planets. *Science*. 2012;337(6095):721–723.
53. Piani L, Marrocchi Y, Rigaudier T, Vacher LG, Thomassin D, Marty B. Earth's water may have been inherited from material similar to enstatite chondrite meteorites. *Science*. 2020;369(6507):1110–1113.
54. Sharp Z, Olson PL. Multi-element constraints on the sources of volatiles to Earth. *Geochim Cosmochim Acta*. 2022;333:124–135.
55. Young ED, Shahar A, Schlichting HE. Earth shaped by primordial H_2 atmospheres. *Nature*. 2023;616:306–311.
56. Marty B. Meteoritic noble gas constraints on the origin of terrestrial volatiles. *Icarus*. 2022;381:Article 115020.
57. Yokochi R, Marty B. A determination of the neon isotopic composition of the deep mantle. *Earth Planet Sci Lett*. 2004;225(1–2):77–88.
58. Williams CD, Mukhopadhyay S. Capture of nebular gases during Earth's accretion is preserved in deep-mantle neon. *Nature*. 2019;565:78–81.

59. Füri E, Marty B. Nitrogen isotope variations in the Solar System. *Nat Geosci*. 2015;8:515–522.
60. O'Neill HSC. The origin of the Moon and the early history of the Earth–A chemical model. Part 2: The Earth. *Geochim Cosmochim Acta*. 1991;55(4):1159–1172.
61. Steller T, Burkhardt C, Yang C, Kleine T. Nucleosynthetic zinc isotope anomalies reveal a dual origin of terrestrial volatiles. *Icarus*. 2022;386:Article 115171.
62. Libourel G, Marty B, Humbert F. Nitrogen solubility in basaltic melt. Part I. effect of oxygen fugacity. *Geochim Cosmochim Acta*. 2003;67(21):4123–4135.
63. Dixon JE, Stolper EM, Holloway JR. An experimental study of water and carbon dioxide solubilities in mid-ocean ridge basaltic liquids. Part I: Calibration and solubility models. *J Petrol*. 1995;36(6):1607–1631.
64. Stolper E. Water in silicate glasses: An infrared spectroscopic study. *Contrib Mineral Petrol*. 1982;81:1–17.
65. Hirschmann M, Withers A, Ardia P, Foley NT. Solubility of molecular hydrogen in silicate melts and consequences for volatile evolution of terrestrial planets. *Earth Planet Sci Lett*. 2012;345–348:38–48.
66. Sossi PA, Tollan PM, Badro J, Bower DJ. Solubility of water in peridotite liquids and the prevalence of steam atmospheres on rocky planets. *Earth Planet Sci Lett*. 2023;601:Article 117894.
67. Newcombe ME, Brett A, Beckett JR, Baker MB, Newman S, Guan Y, Eiler JM, Stolper EM. Solubility of water in lunar basalt at low pH_2O. *Geochim Cosmochim Acta*. 2017;200: 330–352.
68. Bernadou F, Gaillard F, Füri E, Marrocchi Y, Slodczyk A. Nitrogen solubility in basaltic silicate melt - implications for degassing processes. *Chem Geol*. 2021;573:Article 120192.
69. Fincham C, Richardson FD. The behaviour of Sulphur in silicate and aluminate melts. *Proc R Soc A: Math Phys Eng Sci*. 1954;223(1152):40.
70. O'Neill HSC, Mavrogenes JA. The sulfide capacity and the sulfur content at sulfide saturation of silicate melts at 1400 c and 1 bar. *J Petrol*. 2002;43(6):1049–1087.
71. Boulliung J, Wood BJ. So2 solubility and degassing behavior in silicate melts. *Geochim Cosmochim Acta*. 2022;336: 150–164.
72. Sossi PA, Burnham AD, Badro J, Lanzirotti A, Newville M, O'Neill HSC. Redox state of Earth's magma ocean and its Venus-like early atmosphere. *Sci Adv*. 2020;6(48): Article eabd1387.
73. Lichtenberg T, Bower DJ, Hammond M, Boukrouche R, Sanan P, Tsai S-M, Pierrehumbert RT. Vertically resolved magma ocean–protoatmosphere evolution: H_2, H_2O, CO_2, CH_4, CO, O_2, and N_2 as primary absorbers. *J Geophys Res Planets*. 2021;126(2):Article e2020JE006711.
74. Miller SL. A production of amino acids under possible primitive Earth conditions. *Science*. 1953;117(3046):528–529.
75. Chase MW. *NIST-JANAF thermochemical tables. Washington (DC)*: American Chemical Society; 1998.
76. Bower DJ, Hakim K, Sossi PA, Sanan P. Retention of water in terrestrial magma oceans and carbon-rich early atmospheres. *Planet Sci J*. 2022;3(4):93.
77. O'Neill HSC, Pownceby MI. Thermodynamic data from redox reactions at high temperatures. I. An experimental and theoretical assessment of the electrochemical method using stabilized zirconia electrolytes, with revised values for the Fe-"FeO", Co-CoO, Ni-NiO and Cu-Cu_2O oxygen buffers, and new data for the W-WO_2 buffer. *Contrib Mineral Petrol*. 1993;114:296–314.
78. Gaillard F, Bernadou F, Roskosz M, Bouhifd MA, Marrocchi Y, Iacono-Marziano G, Moreira M, Scaillet B, Rogerie G. Redox controls during magma ocean degassing. *Earth Planet Sci Lett*. 2022;577:Article 117255.
79. Boulliung J, Wood BJ. Sulfur oxidation state and solubility in silicate melts. *Contrib Mineral Petrol*. 2023;178:Article 56.
80. Salvador A, Avice G, Breuer D, Gillmann C, Lammer H, Marcq E, Raymond SN, Sakuraba H, Scherf M, Way MJ. Magma ocean, water, and the early atmosphere of venus. *Space Sci Rev*. 2023;219:Article 51.
81. Maurice M, Dasgupta R, Hassanzadeh P. Redox evolution of the crystallizing terrestrial magma ocean and its influence on the outgassed atmosphere. *Planet Sci J*. 2023;4(2):31.
82. Liggins P, Jordan S, Rimmer PB, Shorttle O. Growth and evolution of secondary volcanic atmospheres: I. Identifying the geological character of hot rocky planets. *J Geophys Res Planets*. 2022;127(7):Article e2021JE007123.
83. Zahnle KJ, Lupu R, Catling DC, Wogan N. Creation and evolution of impact-generated reduced atmospheres of early Earth. *Planet Sci J*. 2020;1(1):11.
84. Itcovitz JP, ASP R, Citron RI, Stewart ST, Sinclair CA, Rimmer PB, Shorttle O. Reduced atmospheres of post-impact worlds: The early Earth. *Planet Sci J*. 2022;3:115.
85. Thompson MA, Krissansen-Totton J, Wogan N, Telus M, Fortney JJ. The case and context for atmospheric methane as an exoplanet biosignature. *Proc Natl Acad Sci U S A*. 2022;119(14):Article e2117933119.
86. Wogan NF, Catling DC, Zahnle KJ, Lupu R. Origin of life molecules in the atmosphere after big impacts on the early Earth. ArXiv. 2023. https://doi.org/10.48550/arXiv.2307.09761
87. Schubert G, Turcotte D, Solomon S, Sleep N. Coupled evolution of the atmospheres and interiors of planets and satellites. In: *Origin and evolution of planetary and satellite atmospheres*. Tucson (AZ): The University of Arizona Press; 1989. pp. 450–483.
88. Lenardic A, Jellinek A, Moresi L-N. A climate induced transition in the tectonic style of a terrestrial planet. *Earth Planet Sci Lett*. 2008;271(1–4):34–42.
89. Gillmann C, Tackley P. Atmosphere/mantle coupling and feedbacks on Venus. *J Geophys Res Planets*. 2014;119(6): 1189–1217.
90. Krissansen-Totton J, Fortney JJ, Nimmo F. Was Venus ever habitable? Constraints from a coupled interior–atmosphere–redox evolution model. *Planet Sci J*. 2021;2(5):216.
91. Ortenzi G, Noack L, Sohl F, Guimond CM, Grenfell JL, Dorn C, Schmidt JM, Vulpius S, Katyal N, Kitzmann D, et al. Mantle redox state drives outgassing chemistry and atmospheric composition of rocky planets. *Sci Rep*. 2020;10:Article 10907.
92. Foley BJ, Driscoll PE. Whole planet coupling between climate, mantle, and core: Implications for rocky planet evolution. *Geochem Geophys Geosyst*. 2016;17(5):1885–1914.
93. Wegener A. Die entstehung der kontinente und ozeane the origin of continents and oceans. In: *Sammlung Vieweg*. F. Vieweg & Sohn; 1915. p. 23.
94. McKenzie DP, Parker RL. The North Pacific: An example of tectonics on a sphere. *Nature*. 1967;216:1276–1280.
95. Morgan WJ. Rises, trenches, great faults, and crustal blocks. *J Geophys Res*. 1968;73(6):1959–1982.
96. Pichon XL. Sea-floor spreading and continental drift. *J Geophys Res*. 1968;73(12):3661–3697.
97. Nataf HC, Richter FM. Convection experiments in fluids with highly temperature-dependent viscosity and the thermal

evolution of the planets. *Phys Earth Planet Inter*. 1982;29(3–4):320–329.
98. Christensen UR. Heat transport by variable viscosity convection and implications for the Earth's thermal evolution. *Phys Earth Planet Inter*. 1984;35(4):264–282.
99. Solomatov VS. Scaling of temperature- and stress-dependent viscosity convection. *Phys Fluids*. 1995;7:266–274.
100. Turcotte DL. An episodic hypothesis for Venusian tectonics. *J Geophys Res Planets*. 1993;98(E9):17061–17068.
101. Moresi L, Solomatov V. Mantle convection with a brittle lithosphere: Thoughts on the global tectonic styles of the Earth and Venus. *Geophys J Int*. 1998;133(3):669–682.
102. Stein C, Schmalzl J, Hansen U. The effect of rheological parameters on plate behaviour in a self-consistent model of mantle convection. *Phys Earth Planet Inter*. 2004;142(3–4): 225–255.
103. Armann M, Tackley PJ. Simulating the thermochemical magmatic and tectonic evolution of Venus's mantle and lithosphere: Two-dimensional models. *J Geophys Res Planets*. 2012;117(12):Article E12003.
104. Lourenço DL, Rozel A, Tackley PJ. Melting-induced crustal production helps plate tectonics on Earth-like planets. *Earth Planet Sci Lett*. 2016;439:18–28.
105. Tackley PJ. Self-consistent generation of tectonic plates in time-dependent, three-dimensional mantle convection simulations. *Geochem Geophys Geosyst*. 2000;1(8): Article 1021.
106. Rozel A, Golabek GJ, Näf R. Formation of ridges in a stable lithosphere in mantle convection models with a viscoplastic rheology. *Geophys Res Lett*. 2015;42(12):4770–4777.
107. Rozel A, Golabek GJ, Jain C, Tackley PJ, Gerya T. Continental crust formation on early earth controlled by intrusive magmatism. *Nature*. 2017;545(7654):332–335.
108. Lourenço DL, Rozel AB, Ballmer MD, Tackley PJ. Plutonic-squishy lid: A new global tectonic regime generated by intrusive magmatism on Earth-like planets. *Geochem Geophys Geosyst*. 2020;21(4):Article e08756.
109. Rolf T, Weller M, Gülcher A, Byrne P, O'Rourke JG, Herrick R, Bjonnes E, Davaille A, Ghail R, Gillmann C, et al. Dynamics and evolution of Venus' mantle through time. *Space Sci Rev*. 2022;218:Article 70.
110. Dehant V, Debaille V, Dobos V, Gaillard F, Gillmann C, Goderis S, Grenfell JL, Höning D, Javaux EJ, Karatekin Ö, et al. Geoscience for understanding habitability in the Solar System and beyond. *Space Sci Rev*. 2019;215:Article 42.
111. Sobolev SV, Sobolev AV, Kuzmin DV, Krivolutskaya NA, Petrunin AG, Arndt NT, Radko VA, Vasiliev YR. Linking mantle plumes, large igneous provinces and environmental catastrophes. *Nature*. 2011;477(7364):312–316.
112. Stern RJ. Is plate tectonics needed to evolve technological species on exoplanets? *Geosci Front*. 2016;7:573–580.
113. Zaffos A, Finnegan S, Peters SE. Plate tectonic regulation of global marine animal diversity. *Proc Natl Acad Sci U S A*. 2017;114(22):5653–5658.
114. Stern RJ, Gerya TV, Chapter 13 - Co-Evolution of Life and Plate Tectonics:The Biogeodynamic Perspective on the Mesoproterozoic-Neoproterozoic Transitions, Duarte JC, editor. Dynamics of Plate Tectonics and Mantle Convection, Elsevier; 2023; pp. 295–319; https://doi.org/10.1016/B978-0-323-85733-8.00013-5.
115. Zerkle AL. Biogeodynamics: Bridging the gap between surface and deep Earth processes. *Philos Trans A Math Phys Eng Sci*. 2018;376(2132):Article 20170401.
116. Leprieur F, Descombes P, Gaboriau T, Cowman PF, Parravicini V, Kulbicki M, Melián CJ, de Santana CN, Heine C, Mouillot D, et al. Plate tectonics drive tropical reef biodiversity dynamics. *Nat Commun*. 2016;7:11461.
117. Pellissier L, Heine C, Rosauer DF, Albouy C. Are global hotspots of endemic richness shaped by plate tectonics? *Biol J Linn Soc Lond*. 2017;123(1):247–261.
118. Tarduno JA, Cottrell RD, Bono RK, Rayner N, Davis WJ, Zhou T, Nimmo F, Hofmann A, Jodder J, Ibañez-Mejia M, et al. Hadaean to Palaeoarchaean stagnant-lid tectonics revealed by zircon magnetism. *Nature*. 2023;618:531–536.
119. Lammer H, Zerkle AL, Gebauer S, Tosi N, Noack L, Scherf M, Pilat-Lohinger E, Güdel M, Grenfell JL, Godolt M, et al. Origin and evolution of the atmospheres of early Venus, Earth and Mars. *Astron Astrophys Rev*. 2018;26(1):Article 2.
120. Rosing MT. ^{13}C-depleted carbon microparticles in >3700-Ma sea-floor sedimentary rocks from West Greenland. *Science*. 1999;283(5402):674–676.
121. Valencia D, O'Connell RJ, Sasselov DD. Inevitability of plate tectonics on super-Earths. *Astrophys J*. 2007;670(1):L45.
122. Valencia D, O'Connell RJ. Convection scaling and subduction on Earth and super-Earths. *Earth Planet Sci Lett*. 2009;286(3–4):492–502.
123. Heck HJV, Tackley PJ. Plate tectonics on super-Earths: Equally or more likely than on Earth. *Earth Planet Sci Lett*. 2011;310(3–4):252–261.
124. Tackley PJ, Ammann M, Brodholt JP, Dobson DP, Valencia D. Mantle dynamics in super-Earths: Post-perovskite rheology and self-regulation of viscosity. *Icarus*. 2013;225:50–61.
125. O'Neill C, Jellinek AM, Lenardic A. Conditions for the onset of plate tectonics on terrestrial planets and moons. *Earth Planet Sci Lett*. 2007;261(1–2):20–32.
126. Stein C, Finnenkötter A, Lowman JP, Hansen U. The pressure-weakening effect in super-Earths: Consequences of a decrease in lower mantle viscosity on surface dynamics. *Geophys Res Lett*. 2011;38(21):Article L21201.
127. Gillmann C, Lognonné P, Chassefière E, Moreira M. The present-day atmosphere of Mars: Where does it come from? *Earth Planet Sci Lett*. 2009;277(3–4):384–393.
128. Gillmann C, Lognonné P, Moreira M. Volatiles in the atmosphere of Mars: The effects of volcanism and escape constrained by isotopic data. *Earth Planet Sci Lett*. 2011;303(3–4):299–309.
129. Grott M, Morschhauser A, Breuer D, Hauber E. Volcanic outgassing of CO_2 and H_2O on Mars. *Earth Planet Sci Lett*. 2011;308(3–4):391–400.
130. Leblanc F, Chassefière E, Gillmann C, Breuer D. Mars' atmospheric 40Ar: A tracer for past crustal erosion. *Icarus*. 2012;218(1):561–570.
131. Lammer H, Chassefière E, Karatekin Ö, Morschhauser A, Niles PB, Mousis O, Odert P, Möstl UV, Breuer D, Dehant V, et al. Outgassing history and escape of the martian atmosphere and water inventory. *Space Sci Rev*. 2013;174:113–154.
132. Ramirez RM, Kopparapu R, Zugger ME, Robinson TD, Freedman R, Kasting JF. Warming early Mars with CO_2 and H_2. *Nat Geosci*. 2014;7:59–63.
133. Way MJ, Ernst RE, Scargle JD. Large-scale volcanism and the heat death of terrestrial worlds. *Planet Sci J*. 2022;3(4):92.
134. Berry AJ, Danyushevsky LV, O'Neill HSC, Newville M, Sutton SR. Oxidation state of iron in komatiitic melt inclusions indicates hot Archaean mantle. *Nature*. 2008;455:960–963.

135. Canil D. Vanadium partitioning and the oxidation state of Archaean komatiite magmas. *Nature*. 1997;389:842–845.
136. Canil D. Vanadium in peridotites, mantle redox and tectonic environments: Archean to present. *Earth Planet Sci Lett*. 2002;195(1–2):75–90.
137. Keller CB, Schoene B. Statistical geochemistry reveals disruption in secular lithospheric evolution about 2.5 Gyr ago. *Nature*. 2012;485:490–493.
138. Gaillard F, Scaillet B, Pichavant M, Iacono-Marziano G. The redox geodynamics linking basalts and their mantle sources through space and time. *Chem Geol*. 2015;418:217–233.
139. Aulbach S, Stagno V. Evidence for a reducing Archean ambient mantle and its effects on the carbon cycle. *Geology*. 2016;44(9):751–754.
140. Armstrong K, Frost DJ, McCammon CA, Rubie DC, Boffa Ballaran T. Deep magma ocean formation set the oxidation state of Earth's mantle. *Science*. 2019;365(6456):903–906.
141. Deng J, Du Z, Karki BB, Ghosh DB, Lee KK. A magma ocean origin to divergent redox evolutions of rocky planetary bodies and early atmospheres. *Nat Commun*. 2020;11(1):2007.
142. Stolper DA, Bucholz CE. Neoproterozoic to early Phanerozoic rise in island arc redox state due to deep ocean oxygenation and increased marine sulfate levels. *Proc Natl Acad Sci U S A*. 2019;116(18):8746–8755.
143. Nicklas RW, Puchtel IS, Ash RD. Redox state of the archean mantle: Evidence from V partitioning in 3.5–2.4 Ga komatiites. *Geochim Cosmochim Acta*. 2018;222:447–466.
144. Frost DJ, McCammon CA. The redox state of Earth's mantle. *Annu Rev Earth Planet Sci*. 2008;36:389–420.
145. Gaillard F, Scaillet B. A theoretical framework for volcanic degassing chemistry in a comparative planetology perspective and implications for planetary atmospheres. *Earth Planet Sci Lett*. 2014;403:307–316.
146. Gaillard F, Bouhifd MA, Füri E, Malavergne V, Marrocchi Y, Noack L, Ortenzi G, Roskosz M, Vulpius S. The diverse planetary ingassing/outgassing paths produced over billions of years of magmatic activity. *Space Sci Rev*. 2021;217(1):Article 22.
147. Gronoff G, Arras P, Baraka S, Bell JM, Cessateur G, Cohen O, Curry SM, Drake JJ, Elrod M, Erwin J, et al. Atmospheric escape processes and planetary atmospheric evolution: From misconceptions to challenges. *J Geophys Res Space Phys*. 2021;125(8): e2019JA027639.
148. Wordsworth R, Kreidberg L. Atmospheres of rocky exoplanets. *Annu Rev Astron Astrophys*. 2022;60:159–201.
149. Bond JC, O'Brien DP, Lauretta DS. The compositional diversity of extrasolar terrestrial planets. I. In situ simulations. *Astrophys J*. 2010;715(2):Article 1050.
150. Moriarty J, Madhusudhan N, Fischer D. Chemistry in an evolving protoplanetary disk: Effects on terrestrial planet composition. *Astrophys J*. 2014;787(1):Article 81.
151. Unterborn CT, Kabbes JE, Pigott JS, Reaman DM, Panero WR. The role of carbon in extrasolar planetary geodynamics and habitability. *Astrophys J*. 2014;793(2):Article 124.
152. Hakim K, van den Berg A, Vazan A, Höning D, van Westrenen W, Dominik Carsten. Thermal evolution of rocky exoplanets with a graphite outer shell. *Astron Astrophys*. 2019;630:A152.
153. Lourenço DL, Rozel AB, Gerya T, Tackley PJ. Efficient cooling of rocky planets by intrusive magmatism. *Nat Geosci*. 2018;11:322–327.
154. Noack L, Rivoldini A, Hoolst TV. Volcanism and outgassing of stagnant-lid planets: Implications for the habitable zone. *Phys Earth Planet Inter*. 2017;269:40–57.
155. Dorn C, Noack L, Rozel AB. Outgassing on stagnant-lid super-Earths. *Astron Astrophys*. 2018;614:Article A18.
156. Breuer D, Moore W. Dynamics and thermal history of the terrestrial planets, the Moon, and Io. In: *Treatise on geophysics: Planets and moons.* Elsevier; 2007. p. 299–348.
157. O'Reilly TC, Davies GF. Magma transport of heat on Io: A mechanism allowing a thick lithosphere. *Geophys Res Lett*. 1981;8(4):313–316.
158. Veeder GJ, Matson DL, Johnson TV, Davies AG, Blaney DL. The polar contribution to the heat flow of Io. *Icarus*. 2004;169(1):264–270.
159. Smrekar SE, Davaille A, Sotin C. Venus interior structure and dynamics. *Space Sci Rev*. 2018;214:Article 88.
160. Plesa A-C, Padovan S, Tosi N, Breuer D, Grott M, Wieczorek MA, Spohn T, Smrekar SE, Banerdt WB. The thermal state and interior structure of Mars. *Geophys Res Lett*. 2018;45(22):12198–12209.
161. Stål T, Reading AM, Fuchs S, Halpin JA, Lösing M, Turner RJ. Properties and biases of the global heat flow compilation. *Front Earth Sci*. 2022;10:Article 963525.
162. Gillmann C, Golabek GJ, Raymond SN, Schönbächler M, Tackley PJ, Dehant V, Debaille V. Dry late accretion inferred from Venus's coupled atmosphere and internal evolution. *Nat Geosci*. 2020;13:265–269.
163. Segura TL, Toon OB, Colaprete A, Zahnle K. Environmental effects of large impacts on Mars. *Science*. 2002;298(5600):1977–1980.
164. Turbet M, Gillmann C, Forget F, Baudin B, Palumbo A, Head J, Karatekin O. The environmental effects of very large bolide impacts on early Mars explored with a hierarchy of numerical models. *Icarus*. 2020;335:Article 113419.
165. Gillmann C, Golabek GJ, Tackley PJ. Effect of a single large impact on the coupled atmosphere-interior evolution of Venus. *Icarus*. 2016;268:295–312.
166. West AJ, Galy A, Bickle M. Tectonic and climatic controls on silicate weathering. *Earth Planet Sci Lett*. 2005;235(1–2): 211–228.
167. Plank T, Manning CE. Subducting carbon. *Nature*. 2019;574:343–352.
168. Katayama I, Karato S-I. Effects of water and iron content on the rheological contrast between garnet and olivine. *Phys Earth Planet Inter*. 2008;166(1–2):57–66.
169. Karato S. Water in the evolution of the Earth and other terrestrial planets. *Treatise on Geophysics*. 2015;9:105.
170. Green DH, Hibberson WO, Kovács I, Rosenthal A. Water and its influence on the lithosphere-asthenosphere boundary. *Nature*. 2010;467:448–451.
171. Gerya TV, Connolly JA, Yuen DA. Why is terrestrial subduction one-sided? *Geology*. 2008;36(1):43–46.
172. Crameri F, Tackley P, Meilick I, Gerya T, Kaus B. A free plate surface and weak oceanic crust produce single-sided subduction on Earth. *Geophys Res Lett*. 2012;39(3):Article L03306.
173. Forsyth D, Uyeda S. On the relative importance of the driving forces of plate motion*. *Geophys J Int*. 1975;43(1):163–200.
174. Sobolev SV, Brown M. Surface erosion events controlled the evolution of plate tectonics on Earth. *Nature*. 2019;570:52–57.
175. Goldblatt C, Watson AJ. The runaway greenhouse: Implications for future climate change, geoengineering and planetary atmospheres. *Philos Trans R Soc A Math Phys Eng Sci*. 2012;370(1974):4197–4216.

176. Kodama T, Nitta A, Genda H, Takao Y, O'ishi R, Abe-Ouchi A, Abe Y. Dependence of the onset of the runaway greenhouse effect on the latitudinal surface water distribution of Earth-like planets. *J Geophys Res Planets*. 2018;123(2):559–574.
177. Kodama T, Genda H, O'ishi R, Abe-Ouchi A, Abe Y. Inner edge of habitable zones for Earth-sized planets with various surface water distributions. *J Geophys Res Planets*. 2019;124:2306–2324.
178. Kodama T, Genda H, Leconte J, Abe-Ouchi A. The onset of a globally ice-covered state for a land planet. *J Geophys Res Planets*. 2021;126(12):Article e2021JE006975.
179. Way MJ, Davies HS, Duarte JC, Green J. The climates of Earth's next supercontinent: Effects of tectonics, rotation rate, and insolation. *Geochem Geophys Geosyst*. 2021;22(8):e09983.
180. Zhao Z, Liu Y, Li W, Liu H, Man K. Climate change of over 20° C induced by continental movement on a synchronously rotating exoplanet. *Astrophys J Lett*. 2021;910:L8.
181. Li X, Hu Y, Guo J, Lan J, Lin Q, Bao X, Yuan S, Wei M, Li Z, Man K, et al. A high-resolution climate simulation dataset for the past 540 million years. *Sci Data*. 2022;9:Article 371.
182. Yang J, Ji W, Zeng Y. Transition from eyeball to snowball driven by sea-ice drift on tidally locked terrestrial planets. *Nat Astron*. 2020;4:58–66.
183. Seales J, Lenardic A. Deep water cycling and the multi-stage cooling of the Earth. *Geochem Geophys Geosyst*. 2020;21:e2020GC009106.
184. Zhang S, Li Y, Leng W, Gurnis M. Photoferrotrophic bacteria initiated plate tectonics in the neoarchean. *Geophys Res Lett*. 2023;50(13):e2023GL103553.
185. Tarduno JA, Blackman EG, Mamajek EE. Detecting the oldest geodynamo and attendant shielding from the solar wind: Implications for habitability. *Phys Earth Planet Inter*. 2014;233:68–87.
186. O'Rourke JG, Gillmann C, Tackley P. Prospects for an ancient dynamo and modern crustal remanent magnetism on Venus. *Earth Planet Sci Lett*. 2018;502:46–56.
187. Trønnes RG, Baron MA, Eigenmann KR, Guren MG, Heyn BH, Løken A, Mohn CE. Core formation, mantle differentiation and core-mantle interaction within Earth and the terrestrial planets. *Tectonophysics*. 2019;760:165–198.
188. Labrosse S. Thermal evolution of the core with a high thermal conductivity. *Phys Earth Planet Inter*. 2015;247:36–55.
189. Nimmo F. Thermal and compositional evolution of the core. In: Schubert G, editor. *Treatise on geophysics*. Amsterdam: Elsevier; 2015. p. 201–219.
190. Lay T, Hernlund J, Buffett BA. Core–mantle boundary heat flow. *Nat Geosci*. 2008;1:25–32.
191. JCE I, Lekić V, Durán C, Drilleau M, Kim D, Rivoldini A, Khan A, Samuel H, Antonangeli D, Banerdt WB, et al. First observations of core-transiting seismic phases on Mars. *Proc Natl Acad Sci U S A*. 2023;120(18):Article e2217090120.
192. Dumoulin C, Tobie G, Verhoeven O, Rosenblatt P, Rambaux N. Tidal constraints on the interior of Venus. *J Geophys Res Planets*. 2017;122:1338–1352.
193. Jacobson SA, Rubie DC, Hernlund J, Morbidelli A, Nakajima M. Formation, stratification, and mixing of the cores of Earth and Venus. *Earth Planet Sci Lett*. 2017;474:375–386.
194. Ramstad R, Barabash S. Do intrinsic magnetic fields protect planetary atmospheres from Stellar Winds? Lessons from ion measurements at Mars, Venus, and Earth. *Space Sci Rev*. 2021;217:Article 36.
195. Brain D, Bagenal F, Ma Y-J, Nilsson H, Stenberg Wieser G. Atmospheric escape from unmagnetized bodies. *J Geophys Res Planets*. 2016;121(12):2364–2385.
196. Maggiolo R, Nilsson H, Wieser GS, Slapak R, Lindkvist J, Hamrin M, de Keyser J. Why an intrinsic magnetic field does not protect a planet against atmospheric escape. *Astron Astrophys*. 2018;614:Article L3.
197. Dong C, Huang Z, Lingam M. Role of planetary obliquity in regulating atmospheric escape: G-dwarf versus M-dwarf Earth-like exoplanets. *Astrophys J Lett*. 2019;882(2):Article L16.
198. Airapetian VS, Barnes R, Cohen O, Collinson GA, Danchi WC, Dong CF, Del Genio AD, France K, Garcia-Sage K, Glocer A, et al. Impact of space weather on climate and habitability of terrestrial-type exoplanets. *Int J Astrobiol*. 2020;19:136.
199. Urey HC. *The planets: Their origin and development*. New Haven: Yale University; 1952.
200. T. M. Donahue, J. B. Pollack, Origin and evolution of the atmosphere of Venus. In: *Venus*. Tucson (AZ): University of Arizona Press; 1983. pp. 1003–1036.
201. Wedepohl KH. The composition of the continental crust. *Geochim Cosmochim Acta*. 1995;59(7):1217–1232.
202. Lécuyer C, Simon L, Guyot F. Comparison of carbon, nitrogen and water budgets on Venus and the Earth. *Earth Planet Sci Lett*. 2000;181(1–2):33–40.
203. Hartmann J, Dürr HH, Moosdorf N, Meybeck M, Kempe S. The geochemical composition of the terrestrial surface (without soils) and comparison with the upper continental crust. *Int J Earth Sci*. 2012;101:365–376.
204. Berner RA, Lasaga AC, Garrels RM. The carbonate-silicate geochemical cycle and its effect on atmospheric carbon dioxide over the past 100 million years. *Am J Sci*. 1983;283(7):641–683.
205. Sleep NH, Zahnle K. Carbon dioxide cycling and implications for climate on ancient Earth. *J Geophys Res Planets*. 2001;106(E1):1373–1399.
206. Krissansen-Totton J, Garland R, Irwin P, Catling DC. Detectability of biosignatures in anoxic atmospheres with the James Webb Space Telescope: A TRAPPIST-1e case study. *Astron J*. 2018;156(3):114.
207. Walker JCG, Hays PB, Kasting JF. A negative feedback mechanism for the long-term stabilization of the Earth's surface temperature. *J Geophys Res Planets*. 1981;86:9776–9782.
208. Kump LR, Brantley SL, Arthur MA. Chemical weathering, atmospheric CO_2, and climate. *Annu Rev Earth Planet Sci*. 2000;28:611–667.
209. Zeebe RE, Westbroek P. A simple model for the $CaCO_3$ saturation state of the ocean: The "Strangelove," the "Neritan," and the "Cretan" ocean. *Geochem Geophys Geosyst*. 2003;4(12):Article 1104.
210. Kelemen PB, Manning CE. Reevaluating carbon fluxes in subduction zones, what goes down, mostly comes up. *Proc Natl Acad Sci U S A*. 2015;112(30):E3997–E4006.
211. Catling DC, Kasting JF. *Atmospheric evolution on inhabited and lifeless worlds*. Cambridge (UK): Cambridge University Press; 2017.
212. Sagan C, Mullen G. Earth and Mars: Evolution of atmospheres and surface temperatures. *Science*. 1972;177(4043):52–56.
213. Catling DC, Zahnle KJ. The Archean atmosphere. *Sci Adv*. 2020;6(9):Article eaax1420.
214. Höning D. The impact of life on climate stabilization over different timescales. *Geochem Geophys Geosyst*. 2020;21(9):Article e2020GC009105.

215. Colbourn G, Ridgwell A, Lenton TM. The time scale of the silicate weathering negative feedback on atmospheric CO_2. *Global Biogeochem Cycles*. 2015;29(5):583–596.
216. Maher K, Chamberlain CP. Hydrologic regulation of chemical weathering and the geologic carbon cycle. *Science*. 2014;343(6178):1502–1504.
217. Foley BJ, Smye AJ. Carbon cycling and habitability of earth-sized stagnant lid planets. *Astrobiology*. 2018;18(7):873–896.
218. Höning D, Tosi N, Spohn T. Carbon cycling and interior evolution of water-covered plate tectonics and stagnant-lid planets. *Astron Astrophys*. 2019;627:Article A48.
219. Krissansen-Totton J, Catling DC. Constraining climate sensitivity and continental versus seafloor weathering using an inverse geological carbon cycle model. *Nat Commun*. 2017;8:15423.
220. Winnick MJ, Maher K. Relationships between CO_2, thermodynamic limits on silicate weathering, and the strength of the silicate weathering feedback. *Earth Planet Sci Lett*. 2018;485:111–120.
221. Hakim K, Bower DJ, Tian M, Deitrick R, Auclair-Desrotour P, Kitzmann D, Dorn C, Mezger K, Heng K. Lithologic controls on silicate weathering regimes of temperate planets. *Planet Sci J*. 2021;2(2):49.
222. Graham RJ, Pierrehumbert R. Thermodynamic and energetic limits on continental silicate weathering strongly impact the climate and habitability of wet, rocky worlds. *Astrophys J*. 2020;896(2):115.
223. Ibarra DE, Caves JK, Moon S, Thomas DL, Hartmann J, Chamberlain CP, Kate M. Differential weathering of basaltic and granitic catchments from concentration-discharge relationships. *Geochim Cosmochim Acta*. 2016;190:265–293.
224. Spaargaren RJ, Wang HS, Mojzsis SJ, Ballmer MD, Tackley PJ. Plausible constraints on the range of bulk terrestrial exoplanet compositions in the solar neighborhood. *Astrophys J*. 2023;948(1):Article 53.
225. Putirka KD, Rarick JC. The composition and mineralogy of rocky exoplanets: A survey of >4000 stars from the Hypatia Catalog. *Am Mineral*. 2019;104(6):817–829.
226. Zolotov MY. Gas-solid interactions on Venus and other Solar System bodies. *Rev Mineral Geochem*. 2018;84(1):351–392.
227. Broecker WS, Peng T-H. The role of $CaCO_3$ compensation in the glacial to interglacial atmospheric CO_2 change. *Global Biogeochem Cycles*. 1987;1(1):15–29.
228. Hakim K, Tian M, Bower DJ, Heng K. Diverse carbonates in exoplanet oceans promote the carbon cycle. *Astrophys J Lett*. 2023;942(1):Article L20.
229. Coogan LA, Gillis KM. Evidence that low-temperature oceanic hydrothermal systems play an important role in the silicate-carbonate weathering cycle and long-term climate regulation. *Geochem Geophys Geosyst*. 2013;14(6):1771–1786.
230. Kite ES, Ford EB. Habitability of exoplanet waterworlds. *Astrophys J*. 2018;864(1):Article 75.
231. Yang J, Boué G, Fabrycky DC, Abbot DS. Strong dependence of the inner edge of the habitable zone on planetary rotation rate. *Astrophys J Lett*. 2014;787(1):Article L2.
232. Wolf ET, Toon OB. Delayed onset of runaway and moist greenhouse climates for Earth. *Geophys Res Lett*. 2014;41(1):167–172.
233. Wolf ET, Toon OB. The evolution of habitable climates under the brightening Sun. *J Geophys Res Atmos*. 2015;120(12):5775–5794.
234. Way MJ, del Genio AD, Aleinov I, Clune TL, Kelley M, Kiang NY. Climates of warm Earth-like planets. I. 3D model simulations. *Astrophys J Suppl Ser*. 2018;239(2):Article 24.
235. Simpson GC. *The distribution of terrestrial radiation*. Edward Stanford; 1929.
236. Komabayasi M. Discrete equilibrium temperatures of a hypothetical planet with the atmosphere and the hydrosphere of one component-two phase system under constant solar radiation. *J Meteorol Soc Jpn*. 1967;45(1):137–139.
237. Ingersoll AP. The runaway greenhouse: A history of water on Venus. *J Atmos Sci*. 1969;26(6):1191–1198.
238. Nakajima S, Hayashi Y-Y, Abe Y. A study on the 'runaway greenhouse effect' with a one-dimensional radiative-convective equilibrium model. *J Atmos Sci*. 1992;49(23):2256–2266.
239. Bean JL, Abbot DS, Kempton EMR. A statistical comparative planetology approach to the hunt for habitable exoplanets and life beyond the Solar System. *Astrophys J Lett*. 2017;841(2):Article L24.
240. Höning D, Baumeister P, Grenfell JL, Tosi N, Way MJ. Early habitability and crustal Decarbonation of a stagnant-lid Venus. *J Geophys Res Planets*. 2021;126(10):Article e2021JE006895.
241. Lehmer OR, Catling DC, Krissansen-Totton J. Carbonate-silicate cycle predictions of Earth-like planetary climates and testing the habitable zone concept. *Nat Commun*. 2020;11:Article 6153.
242. Schlecker M, Apai D, Lichtenberg T, Bergsten G, Salvador A, Hardegree-Ullman KK. Bioverse: The habitable zone inner edge discontinuity as an imprint of runaway greenhouse climates on exoplanet demographics. ArXiv. 2023. https://doi.org/10.48550/arXiv.2309.04518
243. Rauer H, Catala C, Aerts C, Appourchaux T, Benz W, Brandeker A, Christensen-Dalsgaard J, Deleuil M, Gizon L, Goupil M-J, et al. The PLATO 2.0 mission. *Exp Astron*. 2014;38(1-2):249–330.
244. Graham RJ, Lichtenberg T, Pierrehumbert RT. CO_2 ocean bistability on terrestrial exoplanets. *J Geophys Res Planets*. 2022;127(10):e2022JE007456.
245. Zahnle K, Arndt N, Cockell C, Halliday A, Nisbet E, Selsis F, Sleep NH. Emergence of a habitable planet. *Space Sci Rev*. 2007;129:35–78.
246. Klein HP. The Viking mission and the search for life on Mars. *Rev Geophys*. 1979;17(7):1655–1662.
247. Carr MH. Water on early Mars. In: *Ciba Foundation Symposium 202-evolution of hydrothermal ecosystems on Earth (and Mars?) evolution of hydrothermal ecosystems on Earth (And Mars?): Ciba Foundation Symposium 202*. Chichester, UK: John Wiley & Sons Ltd.; 2007. p. 249–272.
248. Bibring J-P, Langevin Y, Mustard JF, Poulet F, Arvidson R, Gendrin A, Gondet B, Mangold N, Pinet P, Forget F, et al. Global mineralogical and aqueous mars history derived from Omega/Mars express data. *Science*. 2006;312(5772):400–404.
249. Squyres S. *Roving Mars: Spirit, opportunity, and the exploration of the red planet*. UK: Hachette; 2005.
250. Wordsworth RD. The climate of early Mars. *Annu Rev Earth Planet Sci*. 2016;44:381–408.
251. Bishop JL, Fairén AG, Michalski JR, Gago-Duport L, Baker LL, Velbel MA, Gross C, Rampe EB. Surface clay formation during short-term warmer and wetter conditions on a largely cold ancient Mars. *Nat Astron*. 2018;2:206–213.

252. Kounaves SP, Chaniotakis NA, Chevrier VF, Carrier BL, Folds KE, Hansen VM, KM ME, O'Neil GD, Weber AW. Identification of the perchlorate parent salts at the Phoenix Mars landing site and possible implications. *Icarus*. 2014;232:226–231.
253. Webster CR, Mahaffy PR, Atreya SK, Flesch GJ, Mischna MA, Meslin P-Y, Farley KA, Conrad PG, Christensen LE, Pavlov AA, et al. Mars methane detection and variability at gale crater. *Science*. 2015;347(6220):415–417.
254. Hu R, Bloom AA, Gao P, Miller CE, Yung YL. Hypotheses for near-surface exchange of methane on Mars. *Astrobiology*. 2016;16(7):539–550.
255. Banerdt WB, Smrekar SE, Banfield D, Giardini D, Golombek M, Johnson CL, Lognonné P, Spiga A, Spohn T, Perrin C, et al. Initial results from the InSight mission on Mars. *Nat Geosci*. 2020;13(3):183–189.
256. Mattingly R, May L. Mars sample return as a campaign. In: *2011 Aerospace Conference*. Big Sky (MT): IEEE; 2011. pp. 1–13.
257. Beaty DW, Grady MM, McSween HY, Sefton-Nash E, Carrier BL, Altieri F, Amelin Y, Ammannito E, Anand M, Benning LG, et al. The potential science and engineering value of samples delivered to Earth by Mars sample return: International MSR Objectives and Samples Team (iMOST). *Meteorit Planet Sci*. 2019;54:S3.
258. O'Rourke JG, Wilson CF, Borrelli ME, Byrne PK, Dumoulin C, Ghail R, Gülcher AJP, Jacobson SA, Korablev O, Spohn T, et al. Venus, the planet: Introduction to the evolution of Earth's sister planet. *Space Sci Rev*. 2023;219:10.
259. Way MJ, Del Genio AD. Venusian habitable climate scenarios: Modeling Venus through time and applications to slowly rotating Venus-like exoplanets. *J Geophys Res Planets*. 2020;125(5):e2019JE006276.
260. Turbet M, Bolmont E, Chaverot G, Ehrenreich D, Leconte J, Marcq E. Day–night cloud asymmetry prevents early oceans on Venus but not on Earth. *Nature*. 2021;598(7880):276–280.
261. Limaye SS, Mogul R, Baines KH, Bullock MA, Cockell C, Cutts JA, Gentry DM, Grinspoon DH, Head JW, Jessup K-L, et al. Venus, an astrobiology target. *Astrobiology*. 2021;21(10):1163–1185.
262. Greaves JS, AMS R, Bains W, Rimmer PB, Sagawa H, Clements DL, Seager S, Petkowski JJ, Sousa-Silva C, Ranjan S, et al. Phosphine gas in the cloud decks of Venus. *Nat Astron*. 2021;5:655–664.
263. Trompet L, Robert S, Mahieux A, Schmidt F, Erwin J, Vandaele AC. Phosphine in Venus' atmosphere: Detection attempts and upper limits above the cloud top assessed from the SOIR/VEx spectra. *Astron Astrophys*. 2021;645:L4.
264. Encrenaz T, Greathouse TK, Marcq E, Widemann T, Bézard B, Fouchet T, Giles R, Sagawa H, Greaves J, Sousa-Silva C. A stringent upper limit of the PH_3 abundance at the cloud top of Venus. *Astron Astrophys*. 2020;643:L5.
265. Snellen IAG, Guzman-Ramirez L, Hogerheijde MR, Hygate APS, van der Tak FFS. Re-analysis of the 267 GHz ALMA observations of Venus. No statistically significant detection of phosphine. *Astron Astrophys*. 2020;644:L2.
266. Thompson MA. The statistical reliability of 267-GHz JCMT observations of Venus: No significant evidence for phosphine absorption. *Mon Not R Astron Soc Lett*. 2021;501(1):L18–L22.
267. Villanueva GL, Cordiner M, Irwin PGJ, de Pater I, Butler B, Gurwell M, Milam SN, Nixon CA, Luszcz-Cook SH, Wilson CF, et al. No evidence of phosphine in the atmosphere of Venus from independent analyses. *Nat Astron*. 2021;5: 631–635.
268. Taylor FW, Svedhem H, Head JW. Venus: The atmosphere, climate, surface, interior and near-space environment of an Earth-like planet. *Space Sci Rev*. 2018;214(1):35–36.
269. Byrne PK, Ghail RC, Gilmore MS, Celâl Şengör AM, Klimczak C, Senske DA, Whitten JL, Khawja S, Ernst RE, Solomon SC. Venus tesserae feature layered, folded, and eroded rocks. *Geology*. 2021;49(1):81–85.
270. Whitten J, Gilmore MS, Brossier J, Byrne PK, Knicely JJ, Smrekar SE. Venus tesserae: The importance of Venus tesserae and remaining open questions. *Bull Am Astron Soc*. 2021;53(4):012.
271. Way MJ, Ostberg C, Foley BJ, Gillmann C, Höning D, Lammer H, O'Rourke J, Persson M, Plesa A-C, Salvador A, et al. Synergies between Venus & exoplanetary observations: Venus and its extrasolar siblings. *Space Sci Rev*. 2023;219(1):13.
272. McKay CP, Smith HD. Possibilities for methanogenic life in liquid methane on the surface of Titan. *Icarus*. 2005;178(1):274–276.
273. Chyba CF, Phillips CB. Possible ecosystems and the search for life on Europa. *Proc Natl Acad Sci*. 2001;98(3):801–804.
274. Barge LM, Rodriguez LE. Life on enceladus? It depends on its origin. *Nat Astron*. 2021;5:740–741.
275. Vance S, Bouffard M, Choukroun M, Sotin C. Ganymede's internal structure including thermodynamics of magnesium sulfate oceans in contact with ice. *Planet Space Sci*. 2014;96:62–70.
276. Sephton MA, Waite JH, Brockwell TG. How to detect life on icy moons. *Astrobiology*. 2018;18(7):843–855.
277. Grasset O, Dougherty MK, Coustenis A, Bunce EJ, Erd C, Titov D, Blanc M, Coates A, Drossart P, Fletcher LN, et al. Jupiter icy moons explorer (JUICE): An esa mission to orbit ganymede and to characterise the Jupiter system. *Planet Space Sci*. 2013;78:1–21.
278. Phillips CB, Pappalardo RT. Europa clipper mission concept: Exploring Jupiter's ocean moon. *EOS Trans Am Geophys Union*. 2014;95(20):165–167.
279. Lorenz RD, Turtle EP, Barnes JW, Trainer MG, Adams DS, Hibbard KE, Sheldon CZ, Zacny K, Peplowski PN, Lawrence DJ, et al. Dragonfly: A rotorcraft lander concept for scientific exploration at titan. *J Hopkins APL Tech Dig*. 2018;34(3):374–387.
280. Barnes JW, Turtle EP, Trainer MG, Lorenz RD, MacKenzie SM, Brinckerhoff WB, Cable ML, Ernst CM, Freissinet C, Hand KP, et al. Science goals and objectives for the dragonfly titan rotorcraft relocatable lander. *Planet Sci J*. 2021;2(4):130.
281. Seager S, Deming D. Exoplanet atmospheres. *ARAA*. 2010;48:631–672.
282. Tinetti G, Drossart P, Eccleston P, Hartogh P, Heske A, Leconte J, Micela G, Ollivier M, Pilbratt G, Puig L, et al. A chemical survey of exoplanets with ARIEL. *Exp Astron*. 2018;46(1):135–209.
283. Gillon M, Triaud AHMJ, Demory B-O, Jehin E, Agol E, Deck KM, Lederer SM, de Wit J, Burdanov A, Ingalls JG, et al. Seven temperate terrestrial planets around the nearby ultracool dwarf star TRAPPIST-1. *Nature*. 2017;542:456–460.
284. Greene TP, Bell TJ, Ducrot E, Dyrek A, Lagage P-O, Fortney JJ. Thermal emission from the Earth-sized exoplanet TRAPPIST-1 b using JWST. *Nature*. 2023;618(7963):39–42.
285. Zieba S, Kreidberg L, Ducrot E, Gillon M, Morley C, Schaefer L, Tamburo P, Koll DDB, Lyu X, Acuña L, et al.

No thick carbon dioxide atmosphere on the rocky exoplanet TRAPPIST-1 c. ArXiv. 2023. https://doi.org/10.48550/arXiv.2306.10150

286. Lustig-Yaeger J, Meadows VS, Lincowski AP. The detectability and characterization of the TRAPPIST-1 exoplanet atmospheres with *JWST*. *Astron J*. 2019;158:27.
287. Des Marais DJ, Harwit MO, Jucks KW, Kasting JF, Lin DNC, Lunine JI, Schneider J, Seager S, Traub WA, Woolf NJ. Remote sensing of planetary properties and biosignatures on extrasolar terrestrial planets. *Astrobiology*. 2002;2(2):153–181.
288. Kasper M, Cerpa Urra N, Pathak P, Bonse M, Nousiainen J, Engler B, Heritier CT, Kammerer J, Leveratto S, Rajani C, et al. PCS — A roadmap for exoearth imaging with the ELT. *The Messenger*. 2021;182:38–43.
289. Snellen I, de Kok R, Birkby JL, Brandl B, Brogi M, Keller C, Kenworthy M, Schwarz H, Stuik R, et al. Combining high-dispersion spectroscopy with high contrast imaging: Probing rocky planets around our nearest neighbors. *Astron Astrophys*. 2015;576:A59.
290. The LUVOIR Team, The LUVOIR mission concept study final report. ArXiv. 2019. https://doi.org/10.48550/arXiv.1912.06219
291. Gaudi BS, Seager S, Mennesson B, Kiessling A, Warfield K, Cahoy K, Clarke JT, Domagal-Goldman S, Feinberg L, Guyon O, Kasdin J, et al. The habitable exoplanet observatory (HabEx) mission concept study final report. ArXiv. 2020. https://doi.org/10.48550/arXiv.2001.06683
292. Feng YK, Robinson TD, Fortney JJ, Lupu RE, Marley MS, Lewis NK, Macintosh B, Line MR. Characterizing Earth analogs in reflected light: Atmospheric retrieval studies for future space telescopes. *Astron J*. 2018;155(5):200.
293. Quanz SP, Crossfield I, Meyer MR, Schmalzl E, Held J. Direct detection of exoplanets in the 3-10 μm range with E-ELT/METIS. *Int J Astrobiol*. 2015;14(2):279–289.
294. Bowens R, Meyer MR, Delacroix C, Absil O, van Boekel R, Quanz SP, Shinde M, Kenworthy M, Carlomagno B, de Xivry GO, et al. Estimating the direct imaging exoplanet yield around stars within 6.5 parsecs. *Astron Astrophys*. 2021;653:A8.
295. Quanz SP, Absil O, Benz W, Bonfils X, Berger J-P, Defrère D, van Dishoeck E, Ehrenreich D, Fortney J, Glauser A, et al. Atmospheric characterization of terrestrial exoplanets in the mid-infrared: Biosignatures, habitability, and diversity. *Exp Astron*. 2022;54(2-3):1197–1221.
296. Quanz SP, Ottiger M, Fontanet E, Kammerer J, Menti F, Dannert F, Gheorghe A, Absil O, Airapetian VS, Alei E, et al. Large Interferometer For Exoplanets (LIFE). I. Improved exoplanet detection yield estimates for a large mid-infrared space-interferometer mission. *Astron Astrophys*. 2022;664:A21.
297. Kammerer J, Quanz SP, Dannert F, LIFE Collaboration. Large Interferometer For Exoplanets (LIFE). VI. Detecting rocky exoplanets in the habitable zones of Sun-like stars. *Astron Astrophys*. 2022;668:A52.
298. Line M, Quanz SP, Schwieterman EW, Fortney JJ, Stevenson KB, Greene T, Zellem R, Morley C, Kataria T, Tremblay L, et al. The importance of thermal emission spectroscopy for understanding terrestrial exoplanets. *Bull Am Astron Soc*. 2019;51(3):271.
299. Kitzmann D, Patzer ABC, von Paris P, Godolt M, Rauer H. Clouds in the atmospheres of extrasolar planets. II. Thermal emission spectra of Earth-like planets influenced by low and high-level clouds. *Astron Astrophys*. 2011;531:A62.
300. Konrad BS, Alei E, Quanz SP, Angerhausen D, Carrión-González Ó, Fortney JJ, Grenfell JL, Kitzmann D, Mollière P, Rugheimer S, et al. Large Interferometer For Exoplanets (LIFE). III. Spectral resolution, wavelength range, and sensitivity requirements based on atmospheric retrieval analyses of an exo-Earth. *Astron Astrophys*. 2022;664:A23.
301. Alei E, Konrad BS, Angerhausen D, Grenfell JL, Mollière P, Quanz SP, Rugheimer S, Wunderlich F, LIFE Collaboration. Large Interferometer For Exoplanets (LIFE). V. Diagnostic potential of a mid-infrared space interferometer for studying Earth analogs. *Astron Astrophys*. 2022;665:A106.
302. Schwieterman EW, Kiang NY, Parenteau MN, Harman CE, DasSarma S, Fisher TM, Arney GN, Hartnett HE, Reinhard CT, Olson SL, et al. Exoplanet biosignatures: A review of remotely detectable signs of life. *Astrobiology*. 2018;18:663–708.
303. Grewal DS, Dasgupta R, Farnell A. The speciation of carbon, nitrogen, and water in magma oceans and its effect on volatile partitioning between major reservoirs of the Solar System rocky bodies. *Geochim Cosmochim Acta*. 2020;280:281–301.
304. Lodders K, Principles and perspectives in cosmochemistry. In: Goswami A, Reddy BE, editors. *Astrophysics and space science proceedings*. Heidelberg (Berlin): Springer; 2010. pp. 379–417.
305. Yoshioka T, Nakashima D, Nakamura T, Shcheka S, Keppler H. Carbon solubility in silicate melts in equilibrium with a Co-Co−2 gas phase and graphite. *Geochim Cosmochim Acta*. 2019;259:129–143.
306. Lourenço DL, Rozel AB. The past and the future of plate tectonics and other tectonic regimes. In: Duarte JC, editor. *Dynamics of plate tectonics and mantle convection*. Elsevier; 2023. p. 181–196.
307. Kadoya S, Tajika E. Outer limits of the habitable zones in terms of climate mode and climate evolution of Earth-like planets. *Astrophys J*. 2019;875(1):Article 7.
308. Turbet M. Two examples of how to use observations of terrestrial planets orbiting in temperate orbits around low mass stars to test key concepts of planetary habitability. In: Di Matteo P, Creevey O, Crida A, Kordopatis G, Malzac J, Marquette J-B, N’Diaye M, Venot O, editors. *SF2A-2019: Proceedings of the annual meeting of the French Society of Astronomy and Astrophysics*. 2019. pp. 341–346.