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Subsurface exolife

Manasvi Lingam[1] and Abraham Loeb[2]

[1]Harvard-Smithsonian Center for Astrophysics, 60 Garden St, Cambridge, MA 02138, USA and [2]Institute for Theory and Computation, Harvard University, Cambridge MA 02138, USA

Research Article

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Author for correspondence: Manasvi Lingam, E-mail:manasvi.lingam@cfa.harvard.edu

Abstract

We study the prospects for life on planets with subsurface oceans, and find that a wide range of planets can exist in diverse habitats with ice envelopes of moderate thickness. We quantify the energy sources available to these worlds, the rate of production of prebiotic compounds, and assess their potential for hosting biospheres. Life on these planets is likely to face challenges, which could be overcome through a combination of different mechanisms. We estimate the number of such worlds, and find that they may outnumber rocky planets in the habitable zone of stars by a few orders of magnitude.

Introduction

The concept of the circumstellar habitable zone (HZ), i.e. the region around a host star where liquid water can exist on the surface of a planet with a given atmospheric composition, has a complex history (Gonzalez, 2005). Over the past two decades, since its first modern formulation (Kasting *et al.*, 1993), there has been a tendency in some quarters to conflate the HZ with the broader notion of habitability as pointed out by the likes of Schulze-Makuch and Guinan (2016); Tasker *et al.* (2017); Moore *et al.* (2017).[1] Hence, it is necessary to clearly distinguish between these two concepts and recognize the limitations (and strengths) of the HZ as a signpost for life. In an early treatise on the HZ, Sagan (1996) emphasized the fact that a diverse range of planets (and moons) lying outside the HZ are not precluded from having water or life-as-we-know-it.[2]

If one takes into account the possibility that potentially habitable worlds outside the HZ can exist, a wide range of habitats are feasible (Lammer *et al.*, 2009). Planets and satellites with subsurface oceans are amongst the most commonly studied worlds in terms of their capacity to sustain biospheres. In our own Solar System, Europa and Enceladus fall distinctly under this category and have been widely considered as possible abodes for life (Chyba, 2000; Marion *et al.*, 2003; Parkinson *et al.*, 2008; Spencer *et al.*, 2009; Vance *et al.*, 2016; Waite *et al.*, 2017) since they appear to host many of the necessary ingredients. In addition, some theoretical models appear to indicate that the outer planets of the TRAPPIST-1 system (Gillon *et al.*, 2017) also possess subsurface oceans (Barr *et al.*, 2017). The class of planets and moons with deep subterranean biospheres (Gold, 1992; Sleep, 2012; Michalski *et al.*, 2018) also merits consideration, since it widens the boundaries of the conventional HZ (McMahon *et al.*, 2013; Cockell, 2014).

Hitherto, we have restricted our discussion only to objects (planets, moons and planetoids) around stars (Dyson, 2003; Abramov and Mojzsis, 2011). However, it was pointed out in Stevenson (1999) that free-floating planets with thick atmospheres may exist in interstellar space with the appropriate conditions for surface life. A related proposal was advocated in Abbot and Switzer (2011), who suggested that free-floating potentially habitable Earth-sized planets with subsurface oceans may exist. Looking even further beyond, several authors have discussed the possibility of life based on alternative biochemistry (Bains, 2004; Benner *et al.*, 2004; Schulze-Makuch and Irwin, 2008; Stevenson *et al.*, 2015). Thus, it is evident that life in the Universe has a vast range of niches that it could occupy, and worlds with subsurface oceans under ice envelopes constitute an important category.

Hence, we shall concern ourselves with the likelihood of life-as-we-know-it existing within subsurface oceans henceforth in our analysis. In Section 'Icy worlds: temperature profile and habitats', we present a simple model for the thickness of the ice layer and examine the range of objects that can exist in different environments.[3] Next, we examine the energy sources for prebiotic chemistry on these planets and the potential routes to the origin of life in Section 'Energy sources and paths for abiogenesis'. We discuss the biological potential of these worlds

[1]As per NASA's Astrobiology Strategy: '*Habitability has been defined as the potential of an environment (past or present) to support life of any kind.*"

[2]By 'life-as-we-know-it', we will refer henceforth to organisms which involve carbon-based chemistry, and water constitutes the solvent.

[3]The object, which can be either free-floating or gravitationally bound, could refer to a planet, moon or planetoid, but we shall label it a 'planet' henceforth to simplify our notation.

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in Section 'Ecosystems in planets with subsurface oceans', and determine the rate of biomass production through different avenues. In Section 'Implications for detection and panspermia', we determine the total number of subsurface planets that may exist and delineate some of the consequences for panspermia and detection. We conclude with a summary of our main results in Section 'Conclusions'.

Icy worlds: temperature profile and habitats

First, we will formulate a simple model for the thickness of the ice envelope, and identify certain 'habitats' where icy worlds can exist.

Temperature profile of icy worlds

In our analysis, we shall assume that the object under consideration comprises a surface ice layer with an subsurface ocean situated below. For the sake of simplicity, we do not consider worlds where the outer layers consist of both ice and rock, which are believed to exist on some Solar system satellites (Schubert *et al.*, 2004; Nimmo and Pappalardo, 2016).[4] Our assumption of an ice envelope and a subsurface ocean implicitly assumes that the water content is sufficiently high to enable the existence of these two layers. However, it must be recognized that the water inventory of planets and satellites has been predicted to vary considerably (Raymond *et al.*, 2007; Mulders *et al.*, 2015; Ciesla *et al.*, 2015; Bergin *et al.*, 2015), and hence the depth and existence of the subsurface ocean cannot be estimated *a priori*.

The planet's heat flux is assumed to arise from a combination of radiogenic and primordial (gravitational contraction) heating; on Earth, it is known that the former contributes approximately 50% of the total heat flux (The KamLAND Collaboration, 2011). The effects of tidal heating are assumed to be negligible in our model, unlike Europa and Enceladus (Barr and Showman, 2009; Hussmann *et al.*, 2010; Spencer and Nimmo, 2013; Choblet *et al.*, 2017a) as well as some free-floating satellite–planet systems (Debes and Sigurdsson, 2007), where tidal dissipation is expected to play an important role. In addition, we do not explicitly consider the heating due to serpentinization reactions that is anticipated to be quite significant on small worlds such as Enceladus and Mimas (Malamud and Prialnik, 2013). A schematic figure of the planet has been depicted in Fig. 1.

We shall determine the thermal profile for the outer ice layer by assuming that the heat is transported via conduction, and not through convection. The latter depends on a wide range of properties for rocky and icy planets, such as the mass, ice grain size, availability of water and mantle rheology to name a few (Schubert *et al.*, 2001; Barr and McKinnon, 2007; Baraffe *et al.*, 2014) and the putative existence of bistable behaviour, multiple steady states, mixed heating and temporal evolution only serves to complicate matters further (Lenardic and Crowley, 2012; Korenaga, 2017). In our own Solar System, it has been proposed that non-Newtonian creep mechanisms are responsible for convective shutdown on some dwarf planets and satellites with potential subsurface oceans (McKinnon, 2006), such as Callisto and Pluto. It has also been suggested that the inclusion of

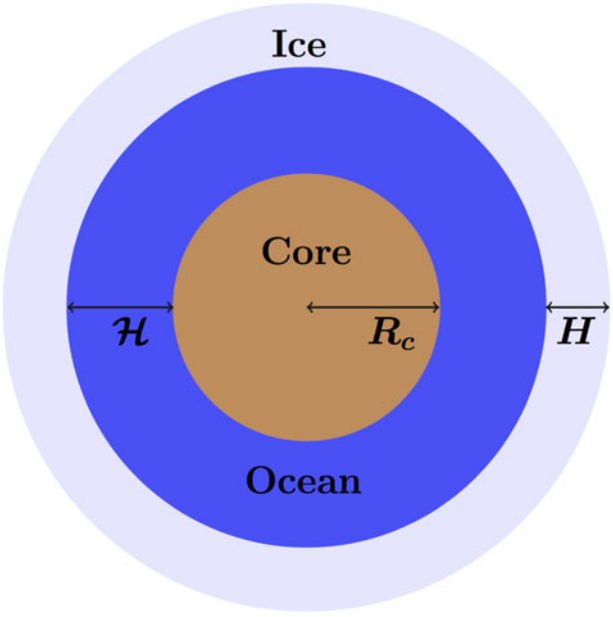


Fig. 1. Schematic illustration of the planet with an ice envelope, a liquid water ocean and an inner rocky/metallic mantle and core.

contaminants (e.g. ammonia and salts), which lower the melting point of ice, may influence the degree of convection (Abbot and Switzer, 2011; Travis *et al.*, 2012).[5] Despite these complexities, provided that the shell's thickness or the heat flow is sufficiently high, thermal convection will occur on such objects, and the thickness of the ice layer can be estimated accordingly (Hussmann *et al.*, 2006; Fu *et al.*, 2010).

Given the above assumptions, the temperature profile is determined via Fourier's law

$$\mathcal{Q} + \kappa \frac{dT}{dr} = 0, \tag{1}$$

where $\mathcal{Q}(r)$ is the geothermal heat flux at radius r (in units of W/m^2), $T \equiv T(r)$ denotes the temperature at radius r, and κ is the thermal conductivity of ice. Note that $\kappa = \mathcal{C}/\mathcal{T}$ with $\mathcal{C} \approx 651$ W/m based on equation (3.11) of Petrenko and Whitworth (1999). The heat flux $\mathcal{Q}(r)$ within the ice envelope can be expressed as

$$\mathcal{Q} = \frac{Q}{4\pi r^2} \times \frac{r^3}{R^3} = \frac{Q\,r}{4\pi R^3}, \tag{2}$$

where R is the radius of the planet, and Q represents the total internal heat flow from the planet to space (in units of W). Thus, the heat flux is given by the ratio of the heat flow within the enclosed region $Q_{\rm enc}$ and the area of this region ($4\pi r^2$). Note that $Q_{\rm enc}$ is approximately equal to $Q \times (\frac{4\pi}{3} r^3 / \frac{4\pi}{3} R^3)$ and the second factor follows from the assumption that the heating sources are uniformly distributed throughout the planet's volume. Furthermore, we will make use of the ansatz

$$Q = \Gamma Q_\oplus \left(\frac{M}{M_\oplus}\right)^\alpha, \tag{3}$$

[4]One such example is Titan, which is believed to possess a subsurface ocean (Iess *et al.*, 2012; Mitri *et al.*, 2014), but we do not consider such moons because of their relatively complex interior structure. These worlds have been predicted to be fairly common around M-dwarfs at distances of $\sim$1 AU (Lunine, 2010).

[5]The presence of clathrate hydrates in the ice shell is predicted to reduce its thickness compared with the pure-ice case for a fixed value of the heat flux (Hand *et al.*, 2006).

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where $Q_\oplus$ is the internal heat flow of the Earth (The KamLAND Collaboration, 2011) and M is the mass of the planet, while Γ and α are free parameters. Note that both Q and $Q_\oplus$ are implicitly time-dependent, as radiogenic heating declines exponentially with time. However, if we restrict the discussion to planets that were formed a few Gyr ago, it can be verified that Q changes merely by an order unity factor (Turcotte and Schubert, 2002) when only long-lived isotopes are taken into consideration. Our analysis has excluded short-lived isotopes (Neveu *et al.*, 2017), but these elements can play a potentially important role in the thermal evolution of icy worlds.

Equation (3) is a generalization of the standard convention wherein $\Gamma = 1$ and $\alpha = 1$. This choice amounts to stating that Q/M, namely the heating rate per unit mass (in W/kg) is constant (Valencia and O'Connell, 2009). For radiogenic heating, Q depends upon the mass of the rocky mantle – usually assumed to have a chondritic composition that contains the radioactive elements (Spohn and Schubert, 2003) – which is not necessarily linearly proportional to the mass of the planet. Hence, there is no *a priori* reason for assuming $\alpha = 1$. Similarly, since planets with metallicities different than that of Earth exist (Johnson and Apps, 2009; Buchhave *et al.*, 2014), it is therefore conceivable that the abundance of radionuclides may also vary accordingly; this variability is encapsulated in our model by means of the parameter Γ.[6] Alternatively, Γ can also be used to encapsulate the degree of heating from other sources (e.g. tidal dissipation).

We turn our attention to the planetary radius which can be determined through a mass–radius relationship. It is not possible to identify a single scaling since it depends on the composition, and is not always a power law (Seager *et al.*, 2007). Nevertheless, for the sake of simplicity, we assume

$$M = \lambda M_\oplus \left(\frac{R}{R_\oplus}\right)^\beta, \tag{4}$$

where R is the planet's radius, with λ and β representing free parameters. The value of β is dependent both on the H_2O content and the mass, but $\lambda \approx 1$ in most cases. We shall use $\beta \approx 3.3$ for $M \lesssim M_\oplus$ (Sotin *et al.*, 2007) and $\beta \approx 3.8$ for $M \gtrsim M_\oplus$ (Valencia *et al.*, 2007; Fu *et al.*, 2010).

Next, we solve (1) by imposing the boundary condition $T(r = R) = T_s$, where T_s denotes the temperature at the surface. The temperature profile is given by

$$\ln\left(\frac{T}{T_s}\right) = \frac{Q(R^2 - r^2)}{8\pi \mathcal{C} R^3}. \tag{5}$$

We are now in a position to determine the thickness of the ice layer. The phase-diagram of H_2O implies that the melting point of ice ranges between ≈ 250 and 270 K provided that the pressure P is lower than 620 MPa (Choukroun and Grasset, 2007). As the temperature dependence is logarithmic in nature, we can assume a melting point of $T_m = 260$ K for pure ice. In contrast, the presence of ammonia (as a contaminant) can lower the melting point of ice to the peritectic temperature of 176 K; this corresponds to a 33% NH_3 concentration in the ocean (Leliwa-Kopystyński *et al.*, 2002). The value of r at which $T = T_m$, denoted by R_m, is

$$R_m = R\left[1 - \frac{8\pi \mathcal{C} R}{Q} \ln\left(\frac{T_m}{T_s}\right)\right]^{1/2}, \tag{6}$$

and the thickness H of the ice layer is determined via $H = R - R_m$. Thus, we obtain

$$\frac{H}{R_\oplus} = \left(\frac{R}{R_\oplus}\right)\left(1 - \left[1 - 2.4 \times 10^{-3} \frac{\ln \Lambda}{\Gamma}\left(\frac{R}{R_\oplus}\right)^{1-\gamma}\right]^{1/2}\right), \tag{7}$$

where we have introduced the auxiliary notation $\gamma = \alpha\beta$ and $\Lambda = T_m/T_s$. By inspecting this formula, it is evident that H will decrease whenever Λ is reduced or Γ is increased for fixed values of $R/R_\oplus$ and γ. Hence, provided that $D > H$, where D denotes the average water depth, the planet can host sub-surface oceans. As noted earlier, it is not possible to quantify D beforehand since the water content of the planet can vary widely.

As an example, let us suppose that the Earth were to be ejected into space. We obtain the surface temperature upon solving $\sigma T_s^4 = \mathcal{F}_\oplus$, where $\mathcal{F}_\oplus = 0.087\ \mathrm{W/m^2}$, and find that $T_s \approx 35$ K. Hence, it follows that $\ln\Lambda \approx 2$ if we use $T_m \approx 260$ K. Using this value in (7) along with $\Gamma = 1$ and $R = R_\oplus$, we find $H \approx 15$ km. The Earth's average water depth $D_\oplus$ is ≈ 3.7 km (Charette and Smith, 2010). Hence, it seems plausible that the Earth's oceans would be entirely frozen as per this model;[7] see also Laughlin and Adams (2000) for a related discussion of this question. If the Earth was ejected shortly after its formation, the geothermal heat flux could have been higher by a factor of order unity due to short-lived radionuclides and primordial heat. Hence, provided that the criterion $\Gamma \gtrsim 4$ was valid, the primordial Earth might have retained a global subsurface ocean if it had become a free-floating planet. For this value of Γ, we obtain $H \approx 3.75$ km for the Earth, and the condition $D_\oplus > H$ is satisfied since $D_\oplus$ for the early Earth was approximately twice its present-day value (Korenaga, 2008).

In Fig. 2, we have plotted H as a function of the radius for different values of the free parameters Γ and Λ. If we consider the case where the second term inside the square brackets of (7) is much smaller than unity, we can use the binomial theorem to obtain the following expression:

$$H \approx 7.6\,\mathrm{km}\ \frac{\ln \Lambda}{\Gamma}\left(\frac{R}{R_\oplus}\right)^{2-\gamma}, \tag{8}$$

and, for $\alpha = 1$ in conjunction with the values of β discussed earlier, the condition $\gamma > 3$ is always satisfied. Hence, the thickness will be a monotonically decreasing function of the radius. Before proceeding further, it must be recognized that (7) is not valid for all values of R. This limitation arises since the quantity inside the square brackets must be positive (to ensure that that its square root is real). Hence, we find that the model has a lower bound (denoted by R_c) given by

$$\frac{R_c}{R_\oplus} = \left(2.4 \times 10^{-3} \frac{\ln \Lambda}{\Gamma}\right)^{1/(\gamma-1)}, \tag{9}$$

[6]In referring to 'metallicity' in this paper, we will work with the astrophysical definition, namely the mass fraction of elements other than hydrogen or helium. The quantity [Fe/H] is often used as a measure of metallicity, and it quantifies the logarithm of the ratio of Fe and H *relative* to the Sun's ratio of Fe and H.

[7]However, even if Earth's oceans were wholly frozen, the possibility of chemolithoautotrophic life in the deep biosphere (e.g. hydrated regions of subduction zones) ought not be discounted (Plümper *et al.*, 2017).

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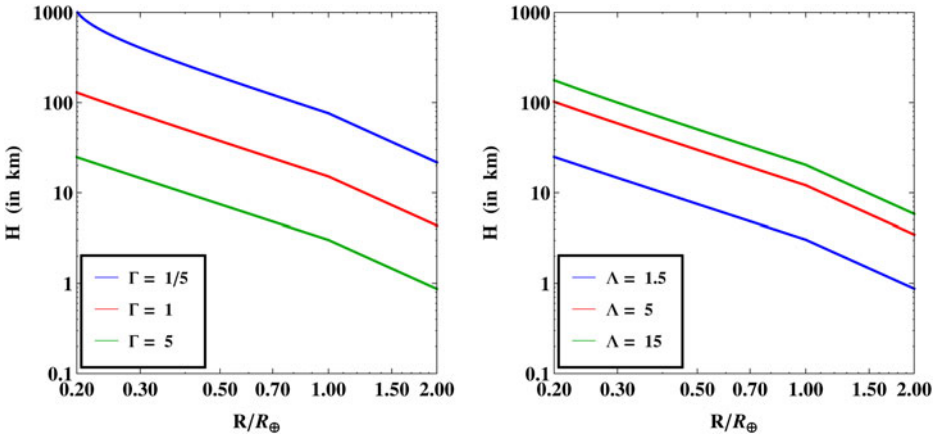


Fig. 2. Left panel: thickness of the ice shell as a function of the planet's radius for differing radionuclide concentrations (represented by Γ) and a fixed surface temperature arising from Earth's radiogenic heating (corresponding to Λ = 7.5). Right panel: thickness of the ice envelope for differing surface temperatures (encapsulated by Λ) and a fixed radionuclide concentration (Γ = 1).

and for the choices $\ln\Lambda \approx 2$ and $\Gamma = 1$, we obtain $R_c \approx 0.1\ R_\oplus$. The value of R_c increases monotonically with Λ and decreases monotonically with Γ. On the other hand, we can assume that $\mathcal{Q}$ is constant in (1) along the lines of Abbot and Switzer (2011). We find that the solution resembles (8) except for the additional factor of Γ and a different power-law exponent.

Our analysis is not expected to yield accurate results for Europa and Enceladus since the effects of tidal dissipation are not included in our model. For instance, if we employ (7) with $\Gamma = 1$ and use $\Lambda \approx 2.5$ for Europa, the resultant value of H is about a factor of 4 higher compared with more sophisticated predictions (Lunine, 2017).

Habitats for potential subsurface oceans

We briefly discuss some of the cases where subsurface oceans might exist.

Type B (involving objects **bound** to a star): When planets lie beyond the HZ of their host star, they can host sub-surface oceans (Ehrenreich and Cassan, 2007; Fu *et al.*, 2010). Examples within our Solar System include Europa and Enceladus, and it has been posited that water-rich extrasolar planets (Léger *et al.*, 2004) such as OGLE 2005-BLG-390Lb may also fall under this category (Ehrenreich *et al.*, 2006). Two important points are worth noting at this stage concerning this class of planets.

- The value of Λ can vary by an order of magnitude because $\Lambda \approx$ 1–10. For instance, the average (morning/evening) surface temperature of Ceres is 168 K (Küppers *et al.*, 2014), which leads to $\Lambda \approx 1.5$ upon choosing $T_m \approx 260$ K; a minimum of $\Lambda \approx 1.1$ can be attained for Ceres at the subsolar point. Recent observations of carbonate deposits in the Occator crater on Ceres (de Sanctis *et al.*, 2016) and the detection of volatiles (Russell *et al.*, 2016; Combe *et al.*, 2016) lend further credence to the possibility that this dwarf planet has/had a subsurface ocean on account of its relatively high surface temperature (Castillo-Rogez and McCord, 2010). On the other hand, for trans-Neptunian objects (TNOs) such as Eris, it can be shown that $\Lambda \approx 10$.
- The HZ need not necessarily correspond only to a main-sequence star. Exoplanets with subsurface oceans (lying outside the HZ) could also exist around white dwarfs and brown dwarfs (Agol, 2011; Barnes and Heller, 2013; Loeb and Maoz, 2013), as well as post-main-sequence (Lorenz *et al.*, 1997; Lopez *et al.*, 2005; Ramirez and Kaltenegger, 2016) and pre-main-sequence (Ramirez and Kaltenegger, 2014; Luger and Barnes, 2015) stars.

If the Type B planet is not very far from its host star, the surface temperature can be estimated via

$$T_s \approx 213\mathrm{K}\left(\frac{1-A}{0.36}\right)^{1/4}\left(\frac{L_\star}{L_\odot}\right)^{1/4}\left(\frac{a}{1\mathrm{AU}}\right)^{-1/2}, \qquad (10)$$

where A is the albedo of the planet, $L_\star$ denotes the luminosity of the host star, and a is the average star–planet distance. We have normalized $1 - A$ by 0.36 since this happens to be the value for Europa. For main-sequence stars, note that $L_\star/L_\odot \propto (M_\star/M_\odot)^\xi$ is commonly used, where $M_\star$ represents the mass of the host star. The value of ξ is dependent on the stellar mass range and is approximately equal to 3.1 for $0.08M_\odot < M_\star < 0.25M_\odot$, 4.5 for $0.25M_\odot < M_\star < 0.75M_\odot$ and 3.5 for $0.75M_\odot < M_\star < 3M_\odot$ (Loeb *et al.*, 2016).

Type U (involving *unbound* objects): When objects are ejected from the planetary system by means of gravitational interactions with giant planets (Rasio and Ford, 1996; Papaloizou and Terquem, 2006), they end up as 'rogue' (i.e. free-floating) planets. Most studies concerning the habitability of these planets have focused on surface-based life-as-we-know-it (Stevenson, 1999; Debes and Sigurdsson, 2007; Badescu, 2011), but it is also possible that they can host subsurface oceans with ice/rock envelopes (Laughlin and Adams, 2000; Abbot and Switzer, 2011).

At first glimpse, it is tempting to conclude that $\Lambda \approx 10$ must hold true in most instances because of the low surface temperature. However, there might exist certain environments where $\Lambda \sim 1$ becomes feasible. We consider one such example in detail, namely galactic nuclei. During their quasar phase, the luminosity of supermassive black holes L_{BH} approximately equals the Eddington luminosity (L_{Edd}) (Kocsis and Loeb, 2014), and we have

$$L_{BH} \approx L_{Edd} = 1.3 \times 10^{37}\mathrm{W}\left(\frac{M_{BH}}{10^6 M_\odot}\right), \qquad (11)$$

where M_{BH} is the mass of the supermassive black hole. We will choose $T_s = 100$ K as a fiducial value, which leads to $\Lambda \approx 2$ when the ice layer has a few per cent ammonia. If we denote the distance from the supermassive black hole by $\mathcal{R}$,

$$\sigma T_s^4 \approx \frac{L_{BH}}{4\pi\mathcal{R}^2}, \qquad (12)$$

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thereby yielding,

$$\mathcal{R} \approx 12\text{pc}\left(\frac{L}{10^{37}\text{W}}\right)^{1/2}\left(\frac{T_\text{s}}{100\text{K}}\right)^{-2}. \tag{13}$$

Although this distance is 2–3 orders of magnitude smaller than the inner edge of the conventional galactic habitable zone (GHZ) which is a few kpc from the Galactic centre (Lineweaver *et al.*, 2004; Forgan *et al.*, 2017) – see, however, Prantzos (2008) and Morrison and Gowanlock (2015) – it must be recognized that the GHZ presupposes the existence of host stars and surficial life. The presence of a sufficiently thick ice layer could potentially shield the planet from ionizing radiation from supernovae, Gamma Ray Bursts as well as other astrophysical catastrophes (Dartnell, 2011; Melott and Thomas, 2011), and effects like atmospheric erosion due to hydrodynamic escape (Forbes and Loeb, 2017; Balbi and Tombesi, 2017) would also be rendered irrelevant.[8] However, at such close distances, it is possible that gravitational interactions play a disruptive role (Genzel *et al.*, 2010). The characteristic quasar lifetime is $\sim 10^7 - 10^8$ years (Martini, 2004), after which the surface temperature would drop and the thickness of the ice envelope would increase by a factor of a few. Nevertheless, even a relatively short interval of time ($\sim 10^7$ yr) might have sufficed for the origin of life on the Earth (Oberbeck and Fogleman, 1989; Lazcano and Miller, 1994; Lineweaver and Davis, 2002) and some worlds may be characterized by similar timescales for the origin of microbial life.[9]

Giant molecular clouds at the Galactic centre (e.g. Sagittarius B2) possess a wide range of organic molecules at relatively high concentrations (Herbst and van Dishoeck, 2009), whose putative significance in prebiotic chemistry has been extensively investigated. Examples include hydrogen cyanide (Jones *et al.*, 2012), aldehydes (Hollis *et al.*, 2000; Requena-Torres *et al.*, 2008), and nitriles (Belloche *et al.*, 2008, 2013).[10] Hence, any type B or U planets existing in these regions may be characterized by $\Lambda \sim 1$ and also have access to these prebiotic molecules, although they will subsequently have to be transported across the ice shell into the ocean.

A second avenue for achieving $\Lambda \sim 1$ is through the cosmic microwave background (CMB). The CMB energy density is redshift dependent, and we determine the surface temperature by equating the CMB energy flux to σT_s^4. Upon doing so, we arrive at

$$T_\text{s} \approx 82\text{K}\left(\frac{1+z}{30}\right) \tag{14}$$

where z is the redshift (Weinberg, 2008). For $z \approx 30$, we obtain $\Lambda \approx 2$ when the ice has contaminants. Hence, planets that formed during this epoch are expected to have relatively thinner ice envelopes and the likelihood of subsurface life originating at this juncture should not be ruled out; our proposal is very akin to the idea that surface life was possible at $z \sim 100$ (Loeb, 2014) when $T_\text{s} \sim 273$ K. A few of the challenges for abiogenesis in the early Universe are described below.

- The first stars must have formed, and seeded the Galaxy with metals through supernovae. The first stars are expected to have formed at $z \lesssim 30$ (Loeb and Furlanetto, 2013; Bromm, 2013), and hence this condition could have been satisfied.
- Although exoplanets have been observed around stars with a wide range of metallicities (Buchhave *et al.*, 2012), the formation of planetesimals has been predicted to depend on the metallicity (Johansen *et al.*, 2009). Hence, the typically low-metallicity environment of the early Universe may have posed difficulties for planet formation (Lineweaver, 2001; Johnson and Li, 2012), although some theoretical models suggest that the first terrestrial planets were capable of forming $\approx$13 Gyr ago (Behroozi and Peeples, 2015; Zackrisson *et al.*, 2016).
- C, H, N, O, P and S are necessary for life-as-we-know-it, implying that they must be available in sufficient quantities. The putative existence of CEMP (carbon-enhanced metal-poor) planets at high redshifts indicates the potential availability of C (Mashian and Loeb, 2016), but the abundance of P and S on high-redshift planets remains poorly constrained.
- The formation of water in molecular clouds, and its subsequent delivery to protoplanetary discs and planets, is an essential requirement (van Dishoeck *et al.*, 2014). Since water vapour is believed to have been abundant even at low metallicity ($\sim 10^{-3}$ of the solar value) in these clouds (Bialy *et al.*, 2015), it suggests that life at high redshifts, only insofar the availability of water is concerned, ought not be ruled out.

If we consider the current epoch, the energy densities of Galactic interstellar radiation, cosmic rays and the CMB are similar (Ferrière, 2001), and collectively yield a surface temperature of a few K; the value of T_s cannot drop below this value. Lastly, for most type U planets, the surface temperature is set by the geothermal heat flux, and is given by

$$T_\text{s} \approx 35\text{K}\Gamma^{1/4}\left(\frac{R}{R_\oplus}\right)^{(\gamma-2)/4}. \tag{15}$$

Energy sources and paths for abiogenesis

Previously, we have seen that free-floating (and bound) planets with subsurface oceans and ice envelopes can exist. However, the availability of liquid water is evidently only a necessary, but *not* sufficient, condition for the planet to be habitable (or inhabited). First, it must be noted that there exist several other factors that must be taken into account in analyses of habitability from a biological standpoint (Cockell *et al.*, 2016). Second, even when liquid water is present, there exist additional stringent constraints set by water activity, chaotropicity, ionic strength, temperature and pressure (Picard and Daniel, 2013; Ball and Hallsworth, 2015; Fox-Powell *et al.*, 2016; Lingam and Loeb, 2018b) since it ought not be regarded merely as a passive background (Ball, 2008; Bellissent-Funel *et al.*, 2016). These reasons collectively serve to explain why the *majority* of Earth's aquasphere (about 88%) is 'not known to host life' (Jones and Lineweaver, 2010).

Hence, it is more instructive to adopt alternative approaches, such as the 'follow the energy' strategy (Hoehler *et al.*, 2007; Shock and Holland, 2007), instead of the 'follow the water' *modus operandi*; the former may also have the advantage of

[8]The erosion rate of Europa's ice layer due to sputtering by energetic ions is $\sim$20 m/Gyr (Cooper *et al.*, 2001). Hence, unless the sputtering rate near the Galactic centre is 2 – 3 orders of magnitude higher, its effects on the ice layer are relatively insignificant over Gyr timescales.

[9]Yet, it is equally important to recognize that a great deal remains unknown about the pathways and timescales for abiogenesis on the Earth and other habitable planets (Orgel, 1998; Spiegel and Turner, 2012).

[10]Kuan *et al.* (2003) claimed to have detected glycine in the Central Molecular Zone, but their evidence does not appear to be supported by subsequent studies (Snyder *et al.*, 2005; Cunningham *et al.*, 2007).

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addressing, to some degree, the prospects for life based on non-standard biochemistries.[11] In this section, we will explore energy considerations and possible routes available for the origin of life (abiogenesis) that must be taken into consideration for assessing the likelihood of these planets to support simple or complex biospheres.

Energy sources for prebiotic synthesis

There remains a great deal that is unknown about the processes that led to abiogenesis on the Earth (Fry, 2000; Ruiz-Mirazo *et al.*, 2014; Luisi, 2016) although it is likely that no single micro-environment or physicochemical process was responsible for the emergence of life on Earth (Delaye and Lazcano, 2005; Spitzer, 2017).[12] Despite these inherent uncertainties, the availability of free energy (Schrödinger, 1944; Deamer, 1997; Dyson, 1999; Pascal *et al.*, 2013; Walker, 2017) is widely regarded as a necessary requirement for abiogenesis to take place. Other basic requirements include: (i) raw materials, (ii) suitable solvent and (iii) appropriate environmental conditions (Hoehler, 2004, 2007). As we focus only on planets where subsurface oceans can exist, (ii) is automatically satisfied. Although (i) and (iii) are undoubtedly important, they are also harder to quantify, and we shall assume these criteria are fulfilled in our subsequent analysis.

On the Earth, ultraviolet (UV) radiation has been identified as one of the most dominant energy sources for enabling prebiotic reactions (Deamer and Weber, 2010). There are several lines of evidence indicating that UV radiation played an important role in prebiotic chemistry on Earth (Patel *et al.*, 2015; Rapf and Vaida, 2016; Sutherland, 2017). By considering the UV radiation with wavelength <200 nm emitted by the host star, we can heuristically evaluate the far-UV (FUV) flux received by the planet in this range (denoted by $\Phi_{\rm UV}$). Upon assuming that the Lyα emission serves as an approximate proxy for FUV radiation, we obtain for type B planets,

$$\Phi_{\rm UV} \sim 10^{6}{\rm Jm}^{-2}{\rm yr}^{-1}\left(\frac{a}{1{\rm AU}}\right)^{-2}\left(\frac{M_{\star}}{M_{\odot}}\right)^{\nu}, \qquad (16)$$

where $\nu \approx 1.2$ for $M_{\star} \lesssim M_{\odot}$ and $\nu \approx 6.8$ for $M_{\odot} < M_{\star} \lesssim 2\,M_{\odot}$ following the scaling relations in Lingam and Loeb (2017c). We note that the power-law exponent is not constant since a significant portion of the FUV emission from low-mass stars (especially M-dwarfs) is from the chromosphere region (France *et al.*, 2013) in contrast to solar-type stars. The above formula implies that larger planets at closer distances around higher-mass stars (but outside the HZ) may be more conducive to prebiotic synthesis. In reality, it must be recognized that $\Phi_{\rm UV}$ also depends on other stellar parameters such as the rotation rate (Linsky *et al.*, 2013).

Several studies have focused on irradiating interstellar ice analogs at low temperatures with UV radiation (Öberg, 2016) – often through a flowing-hydrogen discharge lamp, whose output is divided between the Lyα line and a 20 nm band centred about 160 nm – suggesting that our use of the Lyα proxy in (16) could be a reasonable assumption since a significant fraction of the FUV flux in laboratory experiments comprises of Lyα photons. Some of the organic molecules thus synthesized include amino acids (e.g. alanine, glycine, serine) and the RNA/DNA nucleobases (Muñoz Caro *et al.*, 2002; Elsila *et al.*, 2007; Takano *et al.*, 2007; Nuevo *et al.*, 2008, 2012; Materese *et al.*, 2017). Type U planets, which are free-floating, can traverse through interstellar regions close to O/B-type stars and receive high (but transient) doses of UV radiation leading to the rapid formation of biologically relevant molecules (Throop, 2011). Using (16) and the quantum yield of $\approx 10^{-4}$ for the pathway studied in Muñoz Caro *et al.* (2002), the mass of amino acids ($\mathcal{M}_{\rm UV}$) produced per unit time is

$$\mathcal{M}_{\rm UV} \sim 10^{10}{\rm kg/yr}\left(\frac{R}{R_{\oplus}}\right)^{2}\left(\frac{a}{1{\rm AU}}\right)^{-2}\left(\frac{M_{\star}}{M_{\odot}}\right)^{\nu}. \qquad (17)$$

The production rates of organics determined in the paper ought not be extrapolated *ad infinitum* since they are clearly contingent on the availability of the appropriate reactants,[13] and will also be subject to decomposition by ionizing radiation and other processes. It should also be noted that the above value should be taken with due caution since the quantum yield used is pathway dependent. Finally, laboratory experiments are undertaken in controlled environments endowed with a plentiful supply of the requisite ingredients, implying that our estimates are likely to be upper bounds.

A few caveats regarding UV radiation are in order here. We begin by noting that many of laboratory experiments already presuppose the existence of 'feedstock' organic molecules and typically operate at temperatures that are a few tens of K. There is no guarantee that either these molecules are present, or that the same reactions can function at higher temperatures; for methanol-rich ices, these two factors are likely to be more important than the UV flux and the thickness of the ice monolayers (Öberg *et al.*, 2009). It must also be recognized that UV photolysis can not only facilitate the formation of these compounds but also aid in their decomposition; for instance, at ∼100 K it has been shown that the concentration of amino acids (within the uppermost meter of a pure ice layer) will be halved in a span of ∼10 yrs (Orzechowska *et al.*, 2007). Hence, any organic molecules deposited or formed on the icy surface must be eventually transported to the subsurface ocean.

In addition, some of the pathways studied in the laboratory (mimicking Earth-like conditions) driven by UV irradiation presuppose the existence of liquid water (Saladino *et al.*, 2004; Bada, 2013; McCollom, 2013; Patel *et al.*, 2015), and the latter does not exist on the surface of either type B or U worlds on a long-term basis. However, transient water could exist as a result of tectonic processes, cryovolcanism and impact cratering events (Kargel *et al.*, 2000). With regards to the latter, the total energy required to melt the ice shell completely ($E_{\rm m}$) is

$$E_{\rm m} \approx 1.6 \times 10^{26}{\rm J}\left(\frac{R}{R_{\oplus}}\right)^{2}\left(\frac{H}{1{\rm km}}\right), \qquad (18)$$

which is determined from multiplying the latent heat of fusion for ice with the total mass of the ice shell. By equating (18) with the kinetic energy of the asteroid, we determine find its mass as

[11] In our terminology, life that is not based on carbon and does not involve water as the solvent is taken to be composed of 'non-standard biochemistry'.

[12] Note that life need not have originated on the Earth *in situ*, and could have been transported from elsewhere by means of panspermia, as discussed further in Sec. 5.2.

[13] Insofar interstellar ices are concerned, it has been concluded recently that they are abundantly available to most young planetary systems (Ciesla and Sandford, 2012; Cleeves *et al.*, 2014).

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follows:

$$E_{\mathrm{m}} = \frac{1}{2} M_{\mathrm{a}} (v_{\mathrm{e}}^2 + v_{\infty}^2), \tag{19}$$

where M_{a} is the asteroid mass, $v_{\mathrm{e}} = \sqrt{2GM/R}$ is the escape velocity of the planet, while v_{∞} is determined from Öpik's theory of gravitational encounters, and has typical values of a few km/s (Sloan *et al.*, 2017). For an Earth-sized planet with an ice envelope that is a few kms thick, the required asteroid mass is $\mathcal{O}(10^{18})$ kg. In our Solar System, only a few asteroids (e.g. Vesta and Pallas) exceed this mass. Even if the ice envelope is completely obliterated, organisms in the deep (subsurface) ocean may still be able to avoid extinction, and can undergo rapid evolutionary diversification afterwards in some instances (Alroy, 2008; Grant *et al.*, 2017).

Next, the energy flux from the CMB, which is a function of the redshift, is found to be

$$\Phi_{\mathrm{CMB}} \sim 4.6 \times 10^2 (1+z)^4 \mathrm{Jm}^{-2}\mathrm{yr}^{-1}, \tag{20}$$

where Φ_{CMB} signifies the CMB energy flux. It is evident that the dependence on the redshift is quite strong; for $z \sim 30$, we find that Φ_{CMB} increases by almost six orders of magnitude compared to the present-day value. However, even at such high redshifts, the peak wavelength of the CMB radiation lies in the far-infrared (far-IR). At such low energies, it appears unlikely that this energy could be effectively utilized for enabling prebiotic chemistry. On the other hand, the surface temperature (14) is governed by the CMB radiation, and can therefore facilitate the existence of subsurface oceans at sufficiently high redshifts since H will decrease when the value of z, and therefore T_{s}, is increased.

Other energy sources for prebiotic chemistry include electrical discharges, shock waves from impacts (Martins *et al.*, 2013; Furukawa *et al.*, 2015) and volcanism. We need not consider the first two sources because the corresponding prebiotic synthesis has been shown to occur in the atmosphere (Ruiz-Mirazo *et al.*, 2014), and we have assumed that these planets either lack an atmosphere altogether or possess a very tenuous one. Similarly, estimating the energy flux of cryovolcanism is not straightforward since it remains poorly constrained for icy worlds such as Europa (Fagents, 2003; Sparks *et al.*, 2016). The next source that we consider is radioactivity, whose surficial flux is denoted by Φ_{rad}. Upon utilizing (2) and (3), we end up with

$$\Phi_{\mathrm{rad}} \sim 2.7 \times 10^6 \mathrm{Jm}^{-2}\mathrm{yr}^{-1} \Gamma \left(\frac{R}{R_{\oplus}} \right)^{\gamma - 2}. \tag{21}$$

Several theoretical and experimental analyses suggest that radioactivity, especially in the form of naturally occurring surficial nuclear reactors (Draganić *et al.*, 1983; Adam, 2007, 2016), played a potentially important role in the origin of life on Earth (Albarrán *et al.*, 1987; Garzón and Garzón, 2001; Zagórski, 2003; Parnell, 2004; Adam *et al.*, 2018). An interesting point worth mentioning in this context is the putative existence of organisms on type B and U planets, such as *Desulforudis audaxviator* on Earth (Chivian *et al.*, 2008), that derive their energy from radioactive decay.

Since we are not aware of any *in situ* experiments that have yielded the G-values for the prebiotic synthesis of amino acids through natural radioactivity,[14] we must resort to an indirect strategy for estimating the yields of biomolecules. As the alpha and beta particle decay mechanisms are akin to irradiation by energetic protons and electrons respectively, we will posit that their efficiencies are similar. Several studies have been undertaken in connection with the synthesis of organic compounds by means of energetic particles with energies of KeV–MeV. A wide range of organic molecules such as hydrogen cyanide, aldehydes, formamide, amino acids and nucleosides have been synthesized either directly or after acid hydrolysis (Gerakines *et al.*, 2004; Bennett and Kaiser, 2007; Hudson *et al.*, 2008; Cassidy *et al.*, 2010; Kim and Kaiser, 2011; Saladino *et al.*, 2015); here, we note that the possible importance of formamide in abiogenesis has been extensively studied (Saladino *et al.*, 2012).

Kobayashi *et al.* (1995) irradiated cometary ice analogues at 77 K with 3 MeV protons, and obtained a G-value of $\sim 10^{-4} - 10^{-5}$ for their *specific* setup that yielded amino acids upon hydrolysis. We will work with the smaller value since we anticipate the efficiency of natural radioactivity-mediated synthesis to be lower. Hence, we find that the mass rate of amino acids synthesized ($\mathcal{M}_{\mathrm{rad}}$) is

$$\mathcal{M}_{\mathrm{rad}} \sim 6.8 \times 10^4 \mathrm{kg/yr} \Gamma \left(\frac{R}{R_{\oplus}} \right)^{\gamma - 1} \left(\frac{H}{1\mathrm{km}} \right). \tag{22}$$

An important point worth bearing in mind concerning the prior discussion is that the G-values, which are dependent on the radiation dose, are non-constant (Baratta *et al.*, 2002). Furthermore, as with UV radiation, high-energy particles contribute to both the formation and destruction of organic molecules. Lastly, $\mathcal{M}_{\mathrm{rad}}$ represents an upper bound since we have assumed that all of the energy from radioactive heating in the ice layer, roughly approximated by $\sim Q \times (4\pi R^2 H)/(4\pi R^3/3) \sim Q \times (3H/R)$, is available for prebiotic synthesis. In actuality, amino acids and other organics are likely to be produced through radiolysis only in local environments where radionuclides occur in high concentrations (e.g. natural fission reactors).

Next, we turn our attention to another energy source: energetic particles. As noted previously, they produce a wide variety of biologically relevant compounds. It must be noted that there exist three sources of energetic particles for type B planets, but just one for type U planets. Stellar Energetic Particles (SEPs) and energetic particles emanating from (giant) planetary magnetospheres are unique to type B, while Galactic Cosmic Rays (GCRs) are common to both types B and U. In order for particles from planetary magnetospheres to be a significant energy source, which is expected to be true for Europa (Bolton *et al.*, 2015), it follows that our Type B 'planet' must be a moon.

We will start by determining the flux Φ_{GP} received by a moon orbiting a Jupiter-analogue; if we consider a Saturn-analogue instead, the value of Φ_{GP} becomes much lower and is dependent on complex magnetospheric physics that will not be studied here (Cravens, 2004). From the data provided in Table II of Cooper *et al.* (2001), we obtain

$$\Phi_{\mathrm{GP}} \sim 4 \times 10^6 \mathrm{Jm}^{-2}\mathrm{yr}^{-1} \left(\frac{a_m}{4.5 \times 10^{-3} \mathrm{AU}} \right)^{-2}, \tag{23}$$

[14]The G-value represents the number of molecules formed as products of the chemical reaction when 100 eV of energy has been supplied.

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where a_m is the distance from the moon to the giant planet, the normalization factor 4.5×10^{-3} AU represents the Europa–Jupiter distance, and we have assumed that the energetic particle flux obeys an inverse square-law behaviour (Feynman *et al.*, 1993); note that this assumption works well for the ratio of particle fluxes at Ganymede and Callisto (Cooper *et al.*, 2001). Using the G-values from Kobayashi *et al.* (1995), we find

$$\mathcal{M}_{\rm GP} \sim 1.9 \times 10^8 {\rm kg/yr} \left(\frac{R}{R_\oplus}\right)^2 \left(\frac{a_m}{4.5 \times 10^{-3} {\rm AU}}\right)^{-2}, \quad (24)$$

where $\mathcal{M}_{\rm GP}$ is the mass rate of amino acids synthesized through bombardment of the icy surface by energetic particles from the giant planet's magnetosphere.

It is difficult to evaluate the SEP energy flux for two reasons. First, the physics behind SEPs is complex and the integrated fluence is dependent on the particle acceleration mechanisms and the sites of origin (Reames, 2013). Second, the SEP flux is highly variable since it depends on the stellar age, mass and rotation; it is expected to be significant for low-mass stars with high activity and close-in planets (Lingam *et al.*, 2018; Youngblood *et al.*, 2017), but it may not prove to be a dominant player for middle aged *G*-type stars such as the Sun. We turn our attention to the GCR flux $\Phi_{\rm CR}$ near Earth, which is estimated to be

$$\Phi_{\rm CR} \sim 4.6 \times 10^2 {\rm Jm^{-2} yr^{-1}}, \quad (25)$$

based on the value provided in Kobayashi *et al.* (1998). It can be seen from (20) that the GCR energy flux is approximately equal to the CMB energy flux at $z = 0$. Although this value may appear to be small, it is worth recalling that the cosmic-ray flux increases towards the Galactic centre and, more importantly, constitutes one of the few sources that is universally accessible to type B and U planets. The corresponding mass $\mathcal{M}_{\rm CR}$ of amino acids produced per unit time is

$$\mathcal{M}_{\rm CR} \sim 1.7 \times 10^4 {\rm kg/yr} \left(\frac{R}{R_\oplus}\right)^2. \quad (26)$$

Another important source of prebiotic compounds is the exogenous delivery of organic molecules via interplanetary dust particles (IDPs), comets and meteorites (Pizzarello, 2006; Mumma and Charnley, 2011). From Figure 1 of Chyba and Sagan (1992), it is evident that the delivery rates of organics for IDPs are ~ 3 orders of magnitude greater than comets and ~ 5 orders of magnitude higher than meteorites, although the latter (carbonaceous chondrites in particular) tend to be very rich in organics (Sephton, 2002; Callahan *et al.*, 2011; Pizzarello and Shock, 2017). For Europa, the average organic delivery rate of $\sim 10^3 - 10^4$ kg/yr (Pierazzo and Chyba, 2002) through cometary impacts is approximately consistent with the corresponding rate of $\sim 10^3 - 10^6$ kg/yr for Earth (Chyba and Sagan, 1992). It is also evident that exogenous delivery of prebiotic compounds via comets and meteorites does not apply to type U planets.

Hence, we shall restrict our attention to considering exogenous delivery of organics via IDPs. One crucial point worth mentioning here is that the 'soft landings' of IDPs on the planetary surface has been predicted to require a sufficiently thick atmosphere (Chyba and Phillips, 2002), and it is therefore unclear as to whether IDPs could accumulate on the surface when the atmosphere is either absent or rarefied. However, for the sake of completeness, it is still instructive to estimate the mass of organic molecules delivered by IDPs. Clearly, a universal mass accretion rate for all type B and U planets is not feasible. However, we suggest that the following expression for the mass accretion rate $\dot{M}$ constitutes a reasonable approximation:

$$\dot{M} \approx 4\pi R_{\rm max}^2 \rho_{\rm d} \sigma, \quad (27)$$

where $\sigma = \sqrt{V^2 + c_{\rm s}^2}$, $\rho_{\rm d}$ is the density of the ambient dust particles, $c_{\rm s}$ denotes the sound speed, and V is the relative velocity between the object and the dust. Here, $R_{\rm max} = \max\{R, R_{\rm B}\}$, where $R_{\rm B}$ is the modified Bondi radius (Bondi, 1952) defined as $R_{\rm B} = GM/\sigma^2$. It must be noted here that the accretion of IDPs can occur even when bolides have been decoupled from the gas after the dispersion of the solar nebula. If we consider the scenario where gas is not present, σ should be replaced with V, and in this regime $R_{\rm B}$ becomes the Hoyle–Lyttleton radius (Hoyle and Lyttleton, 1939).

For $R_{\rm max} = R$, we obtain the geometric mass accretion rate. In contrast, for the case $R_{\rm max} = R_{\rm B}$, gravitational focusing leads to the Bondi–Hoyle–Lyttleton accretion rate used in many fields of astronomy (Edgar, 2004). We assume that the accretion rate of organics from IDPs, represented by $\mathcal{M}_{\rm DP}$, is proportional to $\dot{M}$. For Earth, we will choose an organic deposition rate of $\sim 5.7 \times 10^7$ kg/yr; this value is based on the geometric mean of the present-day estimate and the rate at 4.4 Gya that was ~ 3 orders of magnitude higher (Chyba and Sagan, 1992). Consequently, we can express $\mathcal{M}_{\rm DP}$ as

$$\mathcal{M}_{\rm DP} \sim 5.7 \times 10^7 {\rm kg/yr} \left(\frac{R}{R_\oplus}\right)^{2\beta} \left(\frac{\sigma}{26 {\rm km/s}}\right)^{-3} \times \left(\frac{\rho_{\rm d}}{2 \times 10^{-24} {\rm kg/m^3}}\right) \quad (28)$$

when $R_{\rm max} = R_{\rm B}$ holds true. The dust density near the heliosphere is $\sim 2 \times 10^{-24}$ kg/m^3 and the relative inflow velocity (which dominates over the sound speed) of the dust is ~ 26 km/s (Gruen *et al.*, 1994; Krüger *et al.*, 2015). Although the values for σ and ρ_d in the ISM (Mann and Kimura, 2000) can be quite different, we anticipate that the planetary radius will play the most dominant role in governing the magnitude of $\mathcal{M}_{\rm DP}$. In contrast, for the case $R_{\rm max} = R$, we find that $\mathcal{M}_{\rm DP}$ is given by

$$\mathcal{M}_{\rm DP} \sim 5.7 \times 10^7 {\rm kg/yr} \left(\frac{R}{R_\oplus}\right)^2 \left(\frac{\sigma}{26 {\rm km/s}}\right) \times \left(\frac{\rho_{\rm d}}{2 \times 10^{-24} {\rm kg/m^3}}\right). \quad (29)$$

Before moving on, we note that $\mathcal{M}_{\rm DP}$ quantifies the total amount of organics delivered (not just amino acids), and should be viewed as an upper bound.

Lastly, we turn our attention to the abiotic production of amino acids from hydrothermal vents that have attracted much attention as one of the potential sites for abiogenesis (Baross and Hoffman, 1985; Martin *et al.*, 2008; Sojo *et al.*, 2016) through the abiotic synthesis and polymerization of prebiotic compounds (McCollom and Seewald, 2007; Baaske *et al.*, 2007; Budin *et al.*, 2009; Russell *et al.*, 2010; Sousa *et al.*, 2013; Holm *et al.*, 2015). In order for hydrothermal vents to exist, a rock–ocean interface

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is required – in reality, this is a non-trivial condition because sufficiently high pressures at the bottom of the ocean may result in the formation of high-pressure ices (Sohl *et al.*, 2010). As a result, the subsurface ocean would become trapped between two ice layers, which is conventionally expected to have important, but probably negative, ramifications for the habitability of such worlds (Lammer *et al.*, 2009; Noack *et al.*, 2016);[15] we shall not tackle the pressure requirements herein, and will assume henceforth that an ocean–bare rock interface does exist.

Although there are multiple variables involved,[16] we will attempt to quantify the rate of abiotic amino acids produced from alkaline, relatively low-temperature hydrothermal vents (Martin and Russell, 2007; Sousa *et al.*, 2013). Recent observations by the *Cassini-Huygens* mission suggest that this microenvironment is ostensibly present on Enceladus (Hsu *et al.*, 2015; Waite *et al.*, 2017), whose high pH has been interpreted as a consequence of serpentinization through the alteration of ultramafic rocks (Glein *et al.*, 2015). The importance of this process stems from the fact that it serves as the 'mother engine' responsible for the origin of life as per some authors (Russell *et al.*, 2013). In contrast, if the rock–ocean interface is acidic or characterized by a higher temperature, the rate of serpentinization will be significantly altered; furthermore, RNA nucleobases and amino acids have short half-lives at high temperatures and pressures (Levy and Miller, 1998; Aubrey *et al.*, 2009; Kua and Bada, 2011; Lepper *et al.*, 2018). In this regard, we note that it remains controversial as to whether the first lifeforms on the Earth were thermophilic (Akanuma *et al.*, 2013; Weiss *et al.*, 2016) or mesophilic (Miller and Lazcano, 1995; Bada and Lazcano, 2002; Cantine and Fournier, 2018).

Vance *et al.* (2007) used a detailed thermal cracking model for small planets/satellites with oceans and olivine-dominated lithospheres. From Table 2 of that paper, it can be seen that the flux of molecular hydrogen production (H_2) is nearly constant, and ranges between $\sim 10^{13}$ to $\sim 10^{14}$ molecules m^{-2} s^{-1}. Choosing the lower bound, we obtain

$$\mathcal{N}_{H_2} \sim 2.7 \times 10^{11} \mathrm{mol/yr} \left(\frac{R}{R_\oplus} \right)^2, \tag{30}$$

and for Enceladus, we obtain $\mathcal{N}_{H_2} \sim 13$ mol/s, which is nearly equal to the value of ~ 11 mol/s obtained in Table 3 of Steel *et al.* (2017). Equivalently, we obtain $\mathcal{N}_{H_2} \sim 4 \times 10^8$ mol/yr, and it agrees fairly well with the estimate of $\sim 10^9$ mol/yr determined from *Cassini* observations of the Enceladus plume (Waite *et al.*, 2017). For Europa, we arrive at $\mathcal{N}_{H_2} \sim 1.6 \times 10^{10}$ mol/yr, which in very good agreement with the value of $\sim 10^{10}$ mol/yr obtained in Vance *et al.* (2016). For Earth, $\mathcal{N}_{H_2}$ computed from (30) for water–rock interactions is comparable to the production rates of H_2 from the Precambrian continental lithosphere (Lollar *et al.*, 2014). Theoretically, (30) can be understood by adopting the following approach (Vance *et al.*, 2016; Steel *et al.*, 2017):

$$\mathcal{N}_{H_2} = \epsilon V, \tag{31}$$

where V is the volume of the region subject to alterations by serpentinization and ϵ is the conversion factor. V can be further expressed as

$$V = \frac{4\pi}{3} [R_c^3 - (R_c - \langle z \rangle)^3] \approx 4\pi R_c^2 \langle z \rangle, \tag{32}$$

for $\langle z \rangle < R_c$, where R_c denotes the radius of the 'core' region (which comprises of both silicates and metals) and $\langle z \rangle$ denotes the width of the serpentinization front. It might be feasible to approximate it via the root-mean-square diffusion distance, i.e. by using

$$\langle z \rangle \approx \sqrt{2Dt}, \tag{33}$$

with D denoting an effective diffusion constant for the advancement of the serpentinization reaction front, and t is the elapsed time. Moreover, we will use the ansatz $R_c \propto R$ that is known to be valid for terrestrial planets (Valencia *et al.*, 2006). Combining these equations, it can be seen that the scaling $\mathcal{N}_{H_2} \propto R^2$ follows as a result, and is consistent with the estimate provided in (30). However, we wish to caution that our analysis may not be valid for planets larger than the Earth (Vance *et al.*, 2007). Next, we wish to calculate the mass rate of *abiotic* amino acids produced ($\mathcal{M}_{\mathrm{HV}}$) from hydrothermal vents. Using the fact that $\mathcal{M}_{\mathrm{HV}}$ is proportional to $\mathcal{N}_{H_2}$ and the data from Section 2.3 of Steel *et al.* (2017), we arrive at

$$\mathcal{M}_{\mathrm{HV}} \sim 6.7 \times 10^8 \mathrm{kg/yr} \left(\frac{R}{R_\oplus} \right)^2. \tag{34}$$

The abiotic and biotic production rates of amino acids have been claimed to be comparable on Enceladus (Steel *et al.*, 2017), but the latter estimate is contingent on factors such as the existence of methanogens and using adenosine triphosphate (ATP) as a measure of the biomass.

The routes to abiogenesis

Hitherto, our analysis has mostly focused on the *surficial* production of prebiotic compounds, with the exception of radiolysis and hydrothermal vents where the organic molecules would be released into the ocean. In the rest of this section, we shall briefly examine how organics synthesized close to the surface may lead to abiogenesis; the various steps that could have lead to the origin of life from hydrothermal vents have already been documented in detail elsewhere (Russell *et al.*, 2014; Burcar *et al.*, 2015; Kebukawa *et al.*, 2017).

As noted previously, it is necessary for these molecules to penetrate deeper into the surface before they are subject to total decomposition by sputtering, electromagnetic radiation and charged particles. In this context, we observe that gardening (Chyba and Phillips, 2002) in conjunction with tectonics and volcanism (if present) can lead to vertical mixing,[17] and thereby

[15] On the other hand, despite the existence of high-pressure ice, a combination of convection and melting could enable the slow transport of salts and nutrients. These mechanisms, which have been predicted to operate on Ganymede (Choblet *et al.*, 2017b; Kalousová *et al.*, 2018), may collectively offset the challenges posed to long-term habitability.

[16] One such example is the role of water activity (often governed by water–rock interactions) in regulating the rate of serpentinization of olivine; the latter has been observed to decrease in laboratory micro-reactors when the salinity is increased (Lamadrid *et al.*, 2017).

[17] Note that there exists intriguing evidence favouring the presence of both cryovolcanism (Sparks *et al.*, 2017) and subduction (Kattenhorn and Prockter, 2014) on Europa.

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transport the organics to lower regions where they are protected from ionizing radiation (Dartnell, 2011). However, in the case of type U planets, we anticipate that gardening, a non-linear process facilitated due to surface bombardment by micrometeorites, is likely to be absent or minimal.

It is believed that one of the significant challenges faced by prebiotic chemistry is that the appropriate organic compounds must be present in sufficiently high concentrations to undergo chemical reactions (Budin and Szostak, 2010). Second, even at sufficiently high concentrations, these molecules must undergo polymerization to eventually yield peptides and nucleic acids without forming 'tar' (Shapiro, 1984; Benner *et al.*, 2010, 2012). It has been shown that wet–dry cycles and thermal gradients can play an important role in facilitating these processes (Kreysing *et al.*, 2015; Ross and Deamer, 2016). On Earth, a wide range of environments have been identified that fulfill the requisite criteria, such as coastal regions and intermountain valleys (Lingam and Loeb, 2017b). As these environments are not likely to exist on planets with subsurface oceans, it raises a potentially important difficulty.

However, when the flexibility of ice as a medium is taken into account, many of these concerns are alleviated. Several studies have concluded that eutectic freezing serves as an effective mechanism for concentrating prebiotic compounds (Levy *et al.*, 2000; Miyakawa *et al.*, 2002a,b; Bada, 2004; Price, 2007). Laboratory experiments have shown that freeze–thaw cycles, in the presence of suitable catalysts, play a beneficial role in the formation of RNA polymerase ribozymes (Monnard and Szostak, 2008; Attwater *et al.*, 2010, 2013; Mutschler *et al.*, 2015) owing to the stabilizing properties of ice (Bartels-Rausch *et al.*, 2012). However, it is unclear as to whether these cycles are sufficiently important on a global scale since they tend to operate over geologically slow timescales, and do not alter fractional concentrations significantly (for pure ice). Nonetheless, taken collectively, a reasonable case could be built for ice as one of the possible sites for abiogenesis to take place (Trinks *et al.*, 2005). If this hypothesis were indeed valid, the possibility that life may have originated on type B and U planets ought not be ruled out. We have implicitly assumed that the requisite minerals (Lambert, 2008; Hazen and Sverjensky, 2010) and other raw materials are available in sufficient quantities, but this supposition is not always valid (Gaidos *et al.*, 2005).

With regards to the above discussion, it is helpful to further quantify the requirements for polymerization. This can be done by evaluating the conditions under which these reactions become exergonic, i.e. the condition $\Delta G < 0$ must be satisfied, where ΔG denotes the Gibbs free energy of formation (Amend *et al.*, 2013). The corresponding values of ΔG for aqueous and crystalline organic compounds at different temperatures and pressures are well documented (Shock and Helgeson, 1988; LaRowe and Helgeson, 2006). A similar study was undertaken in Kimura and Kitadai (2015), and it was shown (see Fig. 1 of that work) that ΔG is weakly dependent on the pressure, and becomes negative for $T_c \sim 50 - 110$ K based on the choice of polymerization reaction. Assuming that $T_s < 50$ K, we can solve for the depth H_c at which these temperatures are attained by employing (5). Upon solving for H_c, we find that the solution is identical to (7) except for the fact that Λ must be replaced by $\Lambda_c = T_c/T_s$. For planets not much smaller than Earth, the following approximation is valid:

$$\frac{H_c}{H} \approx \frac{\ln \Lambda_c}{\ln \Lambda}. \tag{35}$$

As an illustrative example, let us suppose that we consider a free-floating planet with a geothermal heat flux and radius similar to Earth. For these parameters, we obtain $\ln \Lambda_c \approx 0.36 - 1.15$ and $H_c \approx 2.7 - 8.6$ km. Hence, at a depth of a few km, the formation of peptides would become favourable on thermodynamic grounds. Note that H_c will be lowered as one moves towards the central regions of the Galaxy, where T_s can be higher. Although these reactions are exergonic at H_c, it does not necessarily imply that these reactions take place because their rates are proportional to the Boltzmann factor $\exp\left(-E_a/k_B T\right)$, where E_a is the activation energy and T is the ambient temperature. Hence, given that T_c is much lower than the room temperature, polymerization may occur at very low rates.

Thus, through a combination of the above mechanisms, prebiotic compounds could be concentrated, polymerized and delivered to the subsurface ocean underneath the ice envelope where they can undergo subsequent prebiotic evolution and possibly lead to abiogenesis. An important point worth bearing in mind is that the dependency of the probability of life on the concentration of prebiotic compounds (e.g. amino acids) and nutrients is not well understood (Stribling and Miller, 1987). Based on laboratory experiments, it has been suggested that a monomer concentration of $\sim 0.1 - 1$ mM (mmol/L) would be necessary for initiating prebiotic self-assembly processes (Sanchez *et al.*, 1967; Budin and Szostak, 2010). If we assume that the volume of the planet's ocean is similar to that of Earth, the net delivery rate of amino acids must be $\sim 10^7 - 10^8$ kg/yr to attain these global concentrations over Gyr timescales (in the absence of turnover processes). However, it should be noted that the local synthesis of polymerized biomolecules is considerably enhanced due to the presence of thermodynamic cycles that are governed by microenvironmental factors (Braun *et al.*, 2003; Baaske *et al.*, 2007; Da Silva *et al.*, 2015). Hence, the preceding estimate was based on global considerations, whereas several origin-of-life scenarios posit local regions as the sites of abiogenesis (Deamer, 1997; Stüeken *et al.*, 2013), and the latter environments may possess sufficiently high concentrations of prebiotic molecules and other ingredients (e.g. minerals) that enable life to originate (Ferris, 1993).

As mentioned previously, there is also a possibility that life might have originated within the ice layer. On Earth, psychrophiles have evolved a wide range of genotypic and phenotypic characteristics in order to inhabit a diverse array of sea-ice habitats (Thomas and Dieckmann, 2002; D'Amico *et al.*, 2006; Hodson *et al.*, 2008). In fact, the manifold biological adaptations displayed by polyextremophilc terrestrial sea–ice microbes have led to suggestions that they could survive in some environments on Europa and Enceladus (Greenberg *et al.*, 2000; Marion *et al.*, 2003; Lipps and Rieboldt, 2005; Martin and McMinn, 2018). Hence, these factors suggest that the possibility of certain niches within the sea–ice being occupied by motile microbial organisms ought not be ruled out (Nadeau *et al.*, 2016).

Before proceeding further, we emphasize that the existence of thermodynamically feasible condensation, cyclical processes and polymerization are necessary but not sufficient conditions for life to originate. Hence, although both these processes appear to be likely in the ice envelope, several other requirements – such as the self-assembly of amphiphilic compounds into vesicles (Deamer *et al.*, 2002; Schrum *et al.*, 2010) – must also be fulfilled. It may therefore be possible that no *single* environment contains all of the ingredients necessary for abiogenesis to occur. In that event, the combined action of multiple environments and

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mechanisms could, perhaps, collectively facilitate the origin of life (Hazen, 2017) on these worlds with subsurface oceans.

Ecosystems in planets with subsurface oceans

Next, we shall explore the feasibility of subsurface oceanic ecosystems, and delineate some of the challenges and limitations that they are likely to face. We will restrict ourselves to only aquatic habitats, as sea–ice environments have been briefly explored earlier.

The biological potential of subsurface ecosystems

As subsurface oceans do not have access to sunlight, they are not readily capable of supporting photosynthetic organisms. Furthermore, solar radiation is regarded as the most widely available energy source for life on Earth. Hence, the absence of this pathway has often been invoked to argue that the biological potential of icy moons with subsurface oceans is very low relative to Earth (Reynolds *et al.*, 1983; Jakosky and Shock, 1998; Gaidos *et al.*, 1999; McCollom, 1999; Pascal, 2016). However, several authors have discussed multiple ecosystems that are not dependent on photosynthesis, and we shall examine these possibilities below.

Before proceeding further, one notable point worth bearing in mind is that the water in these subsurface oceans is *not* guaranteed to be habitable. As noted in Section 'Energy sources and paths for abiogenesis', there are multiple constraints imposed by temperature, pH, water activity, etc. that can make the conditions impossible for life-as-we-know-it to exist. For instance, it has been pointed out in Pasek and Greenberg (2012) that oxidants delivered to Europa's oceans from the surface could react with sulphides and the oceans would subsequently undergo acidification, and potentially become inimical to life. Among other factors, a highly acidic ocean may disrupt skeletal biomineralization and induce narcosis (Orr *et al.*, 2005); a lowered pH due to hypercapnia has been identified as one of the putative factors in driving the devastating Permian–Triassic mass extinction event (Knoll *et al.*, 1996, 2007). Hence, if life exists in these oceans, it might have evolved either non-standard biomineralization mechanisms and/or resemble acidophiles on the Earth (Rothschild and Mancinelli, 2001; Baker-Austin and Dopson, 2007).

We start with the putative delivery of organics and oxidants from the icy surface to the subsurface oceans. The importance of oxygen has been thoroughly documented for life on Earth (Lane, 2002), and aerobic metabolism provides about an order of magnitude more energy than anaerobic metabolism for the same quantity of raw materials (Catling *et al.*, 2005; McCollom, 2007; Koch and Britton, 2008). Hence, the availability of oxygen has been identified as a potentially significant rate-limiting step for the evolution of complex (extra)terrestrial life (Mills and Canfield, 2014; Lyons *et al.*, 2014; Chen *et al.*, 2015; Knoll and Nowak, 2017; Catling and Kasting, 2017); the reader should, however, consult Butterfield (2009) and Shields-Zhou and Och (2011) that discuss the subtleties involved. Despite these advantages, it should also be borne in mind that O_2 also has severe deleterious effects: it forms superoxides and peroxides that destroy both enzymes and DNA (Imlay, 2013). In order to cope with these harmful consequences, bacteria have been compelled to evolve suitable adaptions.

It was pointed out in Sec. 3.1 that there is a steady flux of energetic particles to the planetary surface from giant planetary magnetospheres, SEPs and GCRs. Let us denote the total flux, which is the sum of the individual three fluxes, by Φ_{T}. For Type B moons, we anticipate that $\Phi_{\mathrm{T}} \approx \Phi_{\mathrm{GP}}$, while $\Phi_{\mathrm{T}} \approx \Phi_{\mathrm{CR}}$ for Type U planets. The basic setup for the delivery of oxidants to the subsurface ocean is as follows. Ionizing radiation from the aforementioned sources leads to the formation of clathrate hydrates of oxidants (Hand *et al.*, 2006), such as H_2O_2, O_2 and CO_2, through radiolysis on the surface (Johnson *et al.*, 2003). Through a combination of gardening and other geological processes, these compounds are buried in the lower layers and eventually delivered to the subsurface ocean, where they can sustain an indirectly radiation-driven ecosystem (Chyba, 2000; Hand *et al.*, 2006).

As there are multiple steps involved, there is no guarantee that these oxidants would ultimately reach the ocean. Furthermore, there exist several uncertainties regarding the rates of sputtering and gardening. As a result, the estimated rates of O_2 delivered to the Europan ocean have ranged from $\sim 10^5$ mol/yr (Chyba and Phillips, 2001) to $\sim 10^{11}$ mol/yr (Greenberg, 2010) based on the depth of the oxygenation layer and the concentration of surficial radiolytic products. We can estimate the delivery rate of $\mathcal{N}_{O_2}$ by following the approach delineated in Hand *et al.* (2007). Thus, we have

$$\mathcal{N}_{O_2} \sim \frac{4\pi R^2 d_{\mathrm{g}} C_0}{\tau_{\mathrm{d}}}, \tag{36}$$

where d_{g} is the gardening depth, C_0 is the molar concentration (with units mol/m^3) of oxidants and τ_{d} is the delivery time. It appears reasonable to suggest that $C_0 \propto \Phi$ since a higher particle flux leads to more oxidants deposited on the surface, and we use $d_{\mathrm{g}} \propto \tau_{\mathrm{d}}^{1/2}$ (Cooper *et al.*, 2001). Substituting these scaling relations into the above equation and using fiducial values of $\mathcal{N}_{O_2} \sim 10^9$ mol/yr for $\tau_{\mathrm{d}} \sim 50$ Myr for Europa (Hand *et al.*, 2007), we obtain

$$\mathcal{N}_{O_2} \sim 1.7 \times 10^{10} \mathrm{mol/yr} \left(\frac{\tau_{\mathrm{d}}}{50\mathrm{Myr}}\right)^{-1/2} \left(\frac{R}{R_\oplus}\right)^2 \times \left(\frac{a_m}{4.5 \times 10^{-3}\mathrm{AU}}\right)^{-2} \tag{37}$$

and we have made use of (23). In contrast, if we assume that there exist resurfacing processes other than gardening, it is more instructive to introduce the variable $\delta = d_{\mathrm{g}}/\tau_{\mathrm{d}}$ and it equals 4 m/Myr for Europa (Greenberg, 2010), leading to $\mathcal{N}_{O_2} \sim 10^{11}$ mol/yr. Using this data in (36), we end up with

$$\mathcal{N}_{O_2} \sim 4.3 \times 10^{11} \mathrm{mol/yr} \left(\frac{\delta}{1\mathrm{m/Myr}}\right) \left(\frac{R}{R_\oplus}\right)^2 \times \left(\frac{a_{\mathrm{m}}}{4.5 \times 10^{-3}\mathrm{AU}}\right)^{-2}. \tag{38}$$

Next, let us recall that long-lived radionuclides are responsible for generating heat. In addition, they also play an important secondary role: through a combination of alpha, beta and gamma decay processes powered by ^{40}K, ^{232}T, ^{235}U and ^{238}U, the radiolysis of water leads to the formation of O_2 and H_2. A combination of these processes may have led to the production of $\sim 2 \times 10^{10}$ mol/yr of H_2 and $\sim 10^{10}$ mol/yr of O_2 on Earth (Draganić *et al.*, 1991; Draganić, 2005). If we make use of the potentially reasonable assumption that the mass of radioactive isotopes is proportional to the mass of the ocean (Chyba and

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Hand, 2001) and make use of the radionuclide enhancement factor Γ introduced earlier, we obtain the following estimates:

$$\mathcal{N}_{H_2} \sim 5.4 \times 10^9 \text{mol/yr} \Gamma \left(\frac{R}{R_\oplus}\right)^2 \left(\frac{\mathcal{H}}{1\text{km}}\right), \quad (39)$$

$$\mathcal{N}_{O_2} \sim 2.7 \times 10^9 \text{mol/yr} \Gamma \left(\frac{R}{R_\oplus}\right)^2 \left(\frac{\mathcal{H}}{1\text{km}}\right), \quad (40)$$

where $\mathcal{H}$ is the ocean depth, and is distinct from the depth of the ice envelope (H); for the sake of simplicity, we have assumed $H/R \ll 1$ and $\mathcal{H}/R \ll 1$ but these relations are not wholly accurate for small objects like Enceladus. Using the fact that $\mathcal{H} \approx 26 - 31$ km for Enceladus (Thomas *et al.*, 2016), we find $\mathcal{N}_{H_2} \approx 2.5 \times 10^8$ mol/yr. This estimate is in excellent agreement with the value $\sim 1 - 3 \times 10^8$ mol/yr derived by means of a detailed alternative radiolysis model (Bouquet *et al.*, 2017; Waite *et al.*, 2017). Although the overall production rates of O_2 and H_2 are similar, worlds situated in the vicinity of magnetized environments (such as the magnetospheres of giant planets), may eventually develop hemispheric concentration gradients (Chyba and Hand, 2001).

Hence, we have provided two different channels by which H_2 can be synthesized: hydrothermal vents and water radiolysis, whose production rates are given by (30) and (39) respectively. Similarly, we have identified two mechanisms for the production of O_2, namely, the delivery of oxidants from the surface and water radiolysis. The former can be determined from (37) and (38), while the latter is given by (40). As there are several free parameters and processes, it is not easy to identify characteristic values. However, it is clear that the rates of production of oxygen and hydrogen are not very dissimilar, i.e. they differ by 1–2 orders of magnitude in most cases. This feature may imply that an approximate redox balance exists on some of these planets, analogous to Earth and possibly Europa (Vance *et al.*, 2016). It will thus be necessary to take into account the long-term redox history in order to properly assess the habitability of type B and U planets.

By adopting the analysis outlined in Chyba and Phillips (2001) for the energy-limited case, where O_2 is consumed by methanotrophs (Russell *et al.*, 2017), we find

$$\frac{dm_0}{dt} \sim 1.1 \times 10^9 \text{kg/yr} \left(\frac{\mathcal{N}_{O_2}}{10^{10}\text{mol/yr}}\right), \quad (41)$$

with dm_0/dt denoting the rate of production of biomass. Assuming a turnover time of $\sim 10^3$ yr based on studies of Earth's deep biosphere (Hoehler and Jørgensen, 2013),[18] we arrive at the following steady-state biomass (m_c):

$$m_c \sim 1.1 \times 10^{12} \text{kg} \left(\frac{\mathcal{N}_{O_2}}{10^{10}\text{mol/yr}}\right), \quad (42)$$

and the corresponding number of cells and their rate of production (in cells/yr) can be computed from the fact that each cell is $\sim 2 \times 10^{-17}$ kg. In comparison, the global net primary production of the Earth is $\sim 10^{14}$ kg/yr (Field *et al.*, 1998) and the total biomass is $\sim 2 \times 10^{15}$ kg (Landenmark *et al.*, 2015). We can also estimate the biomass produced per year due to reductants instead by using the model proposed in Steel *et al.* (2017) for methanogens.[19] Since the biomass produced is proportional to $\mathcal{N}_{H_2}$, we obtain the production rate

$$\frac{dm_0}{dt} \sim 2 \times 10^7 \text{kg/yr} \left(\frac{\mathcal{N}_{H_2}}{10^{10}\text{mol/yr}}\right), \quad (43)$$

and the steady-state biomass is

$$m_c \sim 2 \times 10^{10} \text{kg} \left(\frac{\mathcal{N}_{H_2}}{10^{10}\text{mol/yr}}\right). \quad (44)$$

In addition to these putative ecosystems, life may derive energy from electrical currents by means of electron-transfer reactions, where the electrons are supplied from the magnetospheres of giant planets, and this process has been dubbed 'direct electrophy' (Stelmach *et al.*, 2018). It should be recognized that this near-surface ecosystem would be feasible only for (type B) moons orbiting giant planets. Using the data tabulated in Figure 2 and Table 1 of Stelmach *et al.* (2018), the maximum steady-state biomass is given by

$$m_c \sim 10^{11} \text{kg} \left(\frac{R}{R_\oplus}\right)^2 \left(\frac{\Phi_e}{10^{10}\text{m}^{-2}\text{s}^{-1}}\right), \quad (45)$$

where Φ_e represents the electron number flux (units of $\text{m}^{-2}\,\text{s}^{-1}$) received at the surface of the moon, assuming that the average energy of the particles is approximately 0.5 MeV.

The steady-state concentration of cells (η) in the ocean can be estimated from m_c via

$$\eta \sim 10^9 \text{cells/m}^3 \left(\frac{m_c}{10^{10}\text{kg}}\right) \left(\frac{R}{R_\oplus}\right)^{-2} \left(\frac{\mathcal{H}}{1\text{km}}\right)^{-1}, \quad (46)$$

and the concentration in the plumes (if they are present) will be a factor of ~ 10 lower. The above value is similar to the concentrations observed in other extreme habitats on the Earth, some of which are delineated below.

- Subglacial lakes and icy environments, such as Vostok and Grímsvötn, are expected to have concentrations of $\sim 10^9 - 10^{10}$ cells/m^3 (Priscu *et al.*, 1999; Price, 2000; Mader *et al.*, 2006; Parnell and McMahon, 2016), although concentrations as low as $\sim 10^6 - 10^7$ cells/m^3 have been identified for Vostok (D'Elia *et al.*, 2008).
- The concentration of microbes in deep granitic rock groundwater is typically $\sim 10^{10} - 10^{12}$ cells/m^3 (Pedersen, 1997; Amend and Teske, 2005; Lin *et al.*, 2006).
- In highly oligotrophic habitats, e.g. the sub-seafloor sediments of the North Pacific Gyre, concentrations of $\sim 10^9$ cells/m^3 exist (Orcutt *et al.*, 2011; Røy *et al.*, 2012).

We conclude by analysis by observing that other forms of life that draw upon alternative sources of energy – such as thermal

[18]However, in extreme (e.g. ice and permafrost) environments at ~ 230 K, the turnover time is predicted to be several orders of magnitude higher (Price and Sowers, 2004).

[19]In this regard, we caution the reader that it is not yet clear as to whether methane may have served as the energy source or a byproduct of early life (Russell and Nitschke, 2017), and the existence of either methanogens or methanotrophs on subsurface worlds is not guaranteed.

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gradients, magnetic fields and gravitation (Schulze-Makuch and Irwin, 2006; Muller and Schulze-Makuch, 2006) – might also exist in type B and U planets, but we shall not analyse them here since their total biomass is probably lower (Schulze-Makuch and Irwin, 2002).

The availability of nutrients

The preceding section indicates that the production of biomass can be quite high, albeit some orders of magnitude lower than Earth, when viewed purely from the perspective of energetics. However, an important point worth reiterating is that life needs more than just an energy source and a solvent. Other criteria include the availability of elements such as boron and phosphorus. The former has been argued to play a key role in bringing about the RNA world (Orgel, 2004; Robertson and Joyce, 2012; Higgs and Lehman, 2015) via the chemical stabilization of ribose (Scorei, 2012). As a result, the access to borates in the Hadean–Archean environment has been posited as a crucial factor in abiogenesis (Grew *et al.*, 2011). Phosphorus plays a vital role in nucleic acids, metabolism (through ATP) and membranes (Westheimer, 1987; Neveu *et al.*, 2013). In addition, trace metals (e.g. molybdenum) play an essential role in bioinorganic chemistry and span mechanisms such as photosynthesis, respiration and DNA synthesis, and even minimal changes in their concentration can lead to pathological effects (Bertini *et al.*, 1994); many of these elements are also closely associated with primary marine productivity and macroevolutionary processes (Anbar and Knoll, 2002; Anbar, 2008).

For certain classes of planets and moons, serious considerations of habitability should take into account biogeochemical cycles (e.g. carbon, sulphur) as well as their likelihood of functioning over geological timescales (Cockell *et al.*, 2016); this approach has also been deemed necessary for understanding the evolution of life on early Earth (Knoll *et al.*, 2016). However, the existence of biogeochemical cycles is *not* a universal requirement for habitability – waterworlds represent an interesting counterexample, as shown recently in Kite and Ford (2018). We will focus on phosphorus herein since it constitutes an essential element in regulating ocean productivity (Föllmi, 1996; Tyrrell, 1999). Phosphorus also plays an important role in cellular and molecular biology as discussed earlier, and has even been referred to as 'the staff of life' in the context of aquatic ecology (Karl, 2000). On Earth, the intimately intertwined evolution of marine productivity and phosphorus availability during the Archean and Proterozoic aeons has been extensively investigated (Bjerrum and Canfield, 2002; Konhauser *et al.*, 2007; Papineau, 2010; Laakso and Schrag, 2014; Kipp and Stüeken, 2017), especially in connection with the apparently coincidental emergence of metazoans and the rise of atmospheric oxygen in the late Neoproterozoic era (Planavsky *et al.*, 2010, 2014; Knoll and Sperling, 2014; Reinhard *et al.*, 2016, 2017; Knoll, 2017) but alternative chronologies have been proposed recently by Mills *et al.* (2014) and Zhang *et al.* (2016).

Our choice of phosphorus is therefore particularly relevant since we are essentially studying (subsurface) ocean planets. Hence, maintaining a balance between phosphorus sources and sinks in the ocean is arguably necessary for a sustaining a reasonably copiotrophic (nutrient-rich) biosphere over geologic timescales. On Earth, the two major conventional sources for phosphorus in the ocean are riverine and atmospheric in nature, and the former is greater than the latter by about an order of magnitude, as seen from Table 1 of Benitez-Nelson (2000). The notable sinks in the ocean include sedimentation of organic material, precipitation leading to phosphorite formation and hydrothermal activity. We shall not concern ourselves with the biological recycling of phosphorus in the ocean, despite its undoubted significance across Earth's geologic history, since this process does not constitute (for the most part) a net sink or source in the current epoch (Schlesinger and Bernhardt, 2013).

It is evident that the two major sources identified above are not likely to be functional for planets that have subsurface oceans. Amongst the sinks, the first two are closely linked with biological processes, and we set them aside since they cannot be easily estimated. In addition, their relative contribution is dependent on redox conditions; for instance, the organic burial of phosphorus is enhanced for oxic sediments (Paytan and McLaughlin, 2007). As a result, we are left with only one mechanism – a sink that would quickly deplete phosphorus from the ocean. In order to compute the amount of phosphorus lost per year, we shall invoke that the assumption that the hydrothermal flux of phosphorus removal is constant, which yields

$$\mathcal{N}_{\mathrm{P}} \sim -3 \times 10^{10} \mathrm{mol/yr} \left(\frac{R}{R_{\oplus}} \right)^2 \tag{47}$$

where $\mathcal{N}_{\mathrm{P}}$ denotes the rate of phosphorus gain/loss. Here, we have assumed that the area of the seafloor is proportional to R^2, and used the fiducial value of $\sim 3 \times 10^{10}$ mol/yr for the overall hydrothermal sink on Earth (Wheat *et al.*, 2003); note that the negative sign in (47) signifies the depletion of phosphorus. If we assume that the total mass of the phosphorus $\mathcal{M}_{\mathrm{P}}$ present in the ocean is proportional to its volume, we have

$$\mathcal{M}_{\mathrm{P}} \sim 8.6 \times 10^{14} \mathrm{mol} \left(\frac{R}{R_{\oplus}} \right)^2 \left(\frac{\mathcal{H}}{1 \mathrm{km}} \right), \tag{48}$$

and the normalization is determined from the fact that Earth has $\sim 3.2 \times 10^{15}$ moles of phosphorus (Benitez-Nelson, 2000). Hence, all of the phosphorus in the ocean will be removed over the timescale τ_{P} given by

$$\tau_{\mathrm{P}} \sim \frac{\mathcal{M}_{\mathrm{P}}}{|\mathcal{N}_{\mathrm{P}}|} \sim 2.9 \times 10^{4} \mathrm{yr} \left(\frac{\mathcal{H}}{1 \mathrm{km}} \right). \tag{49}$$

Hence, the above analysis suggests that the biosphere will be mostly comprised of oligotrophes eventually. However, there exist some possibilities that may mitigate this important issue. We begin by recalling that Earth-like hydrothermal vents are not anticipated to exist on planets with high-pressure ices, where the ocean is sandwiched between the two ice layers. This would exclude the hydrothermal sink mechanism described above. As a result, these worlds may be conducive to hosting biospheres but they would constitute examples of *cycle-independent* habitability – see also Kite and Ford (2018) - distinct from Earth-like planets. From the standpoint of sources, phosphorus may be delivered via cometary impacts and meteorites (Pasek and Lauretta, 2008; Altwegg *et al.*, 2016), and the amount delivered exogenously can be quite high depending on the cratering rate. Under certain circumstances, low-temperature submarine weathering (seawater-basalt interactions) can possibly function as a source although it has been claimed to be a small sink on Earth (Froelich *et al.*, 1982).

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However, there is one process that we have not considered. The complex interplay of resurfacing processes (e.g. ridge formation, dilation) can result in some fraction of the ice envelope being melted into the ocean. This process is partly reminiscent of the deposition of phosphorus into the oceans through ice rafting, and an upper bound of 5×10^{10} mol/yr for this mechanism was proposed in Wallmann (2010); see also Raiswell *et al.* (2006). If we assume that an ice layer of thickness $d_{\rm i}$ melts into the ocean in a turnover time of $\tau_{\rm i}$, we introduce the variable $\delta_{\rm i} = d_{\rm i}/\tau_{\rm i}$ with units of m/Myr. We assume that the concentration of phosphorus in the ice layer is $\mathcal{C}_{\rm P}$ (in units of mmol/kg), which may be supplied through meteorite and comet impacts. With these assumptions, we obtain

$$\mathcal{N}_{\rm P} \sim 4.7 \times 10^{8} \text{mol/yr} \left(\frac{R}{R_{\oplus}}\right)^{2} \left(\frac{\delta_i}{1\text{m/Myr}}\right) \left(\frac{\mathcal{C}_{\rm P}}{1\text{mmol/kg}}\right), \quad (50)$$

and this can easily become comparable to (47) when $\delta_{\rm i}$ and/or $\mathcal{C}_P$ are sufficiently high. For example, if we choose $\mathcal{C}_{\rm P} \sim 18$ mmol/kg based on the composition of Earth's continental crust rocks (Faure, 1998) and $\delta_{\rm i} \sim 4$ m/Myr for Europa (Greenberg, 2010), we find that (50) is approximately equal to (47) for arbitrary values of R. However, we caution that (50) constitutes an upper bound since all of the phosphorus in the melted ice will not be accessible to organisms. More specifically, the phosphorus must be available in the form of chemical compounds that are soluble and active in liquid water, and can therefore be readily used by biota. Furthermore, in equilibrium, the amount of ice melted should be replenished by an equal amount of ice formed via freezing. Hence, it is not clear as to whether (50) would ultimately serve as a net source or sink. Lastly, we note that the potential source mechanism (50) is not expected to be valid for planets with surface oceans that also possess negligible subaerial ice coverage, suggesting that these worlds are relatively likely to have oligotrophic biospheres.[20]

Looking beyond phosphorus, the importance of sulphur in biogeochemistry has been well documented (Falkowski *et al.*, 2008; Schlesinger and Bernhardt, 2013), and its possible role as an energy source for chemoautotrophs in the iron–sulphur world (Wachtershauser, 1990) has also received much attention. Hence, in this context, we note that the role of sulphur cycling has been extensively investigated for Europa (Schulze-Makuch and Irwin, 2008). The concentration of sulphates in the ocean layer may approach or exceed that of Earth, although lower values cannot be ruled out as seen from Table 1 of McKinnon and Zolensky (2003); this model was also used to conclude that the likelihood of massive sulphur beds (preventing life near hydrothermal vents) is not negligible. Moreover, since the chemical energy available for sulphate-reducing lifeforms decreases when the pH is lowered (Zolotov and Shock, 2003), it seems plausible that an acidic ocean with pH $\sim$2.6 (Pasek and Greenberg, 2012), would be unfavourable for these organisms; the pH value of $\sim$2.6 was obtained by calculating the equilibrium molar concentrations based on the oxidant delivery rates proposed in Greenberg (2010). Acidic environments could also possess challenges for the polymerization of biomolecules and the origin of life (Deamer *et al.*, 2006). As with the phosphorus cycle, future studies of subsurface oceans should investigate the oceanic source–sink mechanisms for sulphur and the likelihood of sustaining this biogeochemical cycle over geological timescales (Zolotov and Shock, 2004).

The transitions in evolution

Although the 'biological complexity' has increased over time (Carroll, 2001), much remains unknown about the mechanisms responsible for this process, notwithstanding the manifold recent developments in this area (Adami, 2002; McShea and Brandon, 2010; Koonin, 2011; Goldenfeld and Woese, 2011). Over the past two decades, motivated by the seminal work of Smith and Szathmáry (1995), the approach of modelling evolution as a series of 'evolutionary transitions' has proven to be valuable in understanding how smaller units agglomerate to form larger structures that are subsequently acted upon by natural selection (Okasha, 2006; Calcott and Sterelny, 2011; O'Malley and Powell, 2016; Smith and Morowitz, 2016). The common features underlying these transitions stem from the organization, storage and transmittance of information (Szathmáry and Smith, 1997; Woese, 2004; Jablonka and Lamb, 2014; Davies and Walker, 2016), and the transitions are characterized by increasing complexity, division of labour, and innovations in heredity to name a few (Szathmáry, 2015). Examples of these transitions, not all of which are universally accepted, include the origin of prokaryotes, multicellularity, eukaryotes and eusociality (Szathmáry and Smith, 1995).[21]

We observe that related theoretical frameworks, also entailing the evolution of new species that are increasingly complex, have been explored by several authors, e.g. the *megatrajectories* of Knoll and Bambach (2000), the *singularities* of De Duve (2005), and the *energy expansions* of Judson (2017). The likes of Bieri (1964); Morris (2011); Bogonovich (2011); Rospars (2013); Levin *et al.* (2017); Schulze-Makuch and Bains (2017) have also drawn upon similar approaches to arrive at some general predictions concerning the nature of exo-evolutionary transitions and the likelihood of complex extraterrestrial life.[22] We propose that the adoption of the above methodologies could play an important heuristic role in biological analyses of habitability, since they may enable us to understand the likelihood of these transitions on exoplanets. We will offer a few select examples, in the context of type B and U planets, to qualitatively illustrate how this methodology can be employed.

If we approach evolution from the viewpoint of energy expansions, we find that epoch II, corresponding to anaerobic photosynthesis, is unlikely on planets with subsurface oceans, except for putative ecological niches close to the surface. Naively, one may therefore suggest that epoch III, i.e. the rise of oxygen, is

[20]If the rise of oxygen in Earth's atmosphere was due to the growth/oxidation of continents (Kasting, 2013; Lee *et al.*, 2016) or changes in subaerial volcanism (Kump and Barley, 2007; Holland, 2009; Gaillard *et al.*, 2011), it ought not be easy for oxygen levels to attain sufficiently high concentrations (only insofar these specific mechanisms are concerned) in order for complex life to arise on planets with deep surface oceans. The greater mass of liquid water should also result in the dilution of nutrients and prebiotic molecules, although, at local scales, it may still be feasible to achieve sufficiently high concentrations ostensibly necessary for prebiotic self-assembly and abiogenesis.

[21]The number of these transitions that led to noogenesis (the emergence of intelligence) on Earth remains unsettled, but has been suggested to lie between 4 and 7 by several authors (Carter, 1983; Watson, 2008; Carter, 2008; McCabe and Lucas, 2010); see also: http://mason.gmu.edu/ rhanson/hardstep.pdf

[22]The question as to whether these transitions are 'universal' is indubitably an important one, and depends on the nature and degree of universality of the evolutionary process itself (Williams and Fraústo Da Silva, 2003), and the interplay of contingency (Monod, 1971; Gould, 1989; Mayr, 2001; Gould, 2002; Blount *et al.*, 2008) and convergence (Morris, 2003; Vermeij, 2006; Losos, 2011; Rosenblum *et al.*, 2014).

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not feasible. However, as we have seen in Sec. 4.1, there exist two potential mechanisms for the supply of O_2 on Europa and other planets/moons in similar environments: radiolysis of water, and the delivery of oxidants from the surface (also via radiolysis). The likelihood of epoch IV, which is essentially the manifestation of phagocytosis, cannot either be dismissed or validated *a priori*. However, epoch V, where fire functions as a global energy source, is again unlikely on these planets because not all of the requisite basic ingredients are anticipated to be prevalent.

We now turn our attention to megatrajectories (Knoll and Bambach, 2000), some of which also fall under the category of orthodox evolutionary transitions. When dealing with the earliest stages, e.g. the step from abiogenesis to the Last Universal Common Ancestor, arriving at unambiguous conclusions beyond the identification of the manifold energy sources for life is difficult. Next, we observe that many, but not all, eukaryotes on Earth are reliant on aerobic metabolism (Knoll, 2014, 2015). The issue of oxygen generation has already been discussed in Sec. 4.1 but the likelihood of eukaryogenesis cannot be addressed here in further detail. The factors responsible for the origin of eukaryotes have not yet been conclusively identified, although serial/singular endosymbiosis is expected to have played an important role in multiple respects (Sagan, 1967; Kutschera and Niklas, 2005; Embley and Martin, 2006; de Duve, 2007; Martin *et al.*, 2015; Lane, 2017). Hence, the dependency of eukaryogenesis on specific environmental constraints prevalent in subsurface environments cannot be predicted based on our current knowledge.

When we consider the higher evolutionary transitions or megatrajectories, it may be relatively easier to identify the feasibility of these steps. For instance, eusociality (a 'classical' evolutionary transition) is predominantly terrestrial, and one recent study has tentatively identified the role of nesting as being a precondition for eusociality that is more pronounced on land than on sea (Ruxton *et al.*, 2014). Hence, it seems plausible to some degree, at least for life-as-we-know-it, that not many species would evolve this feature on exoplanets with only oceans. From a related standpoint, land-dwelling organisms have been identified as one of the six megatrajectories since they exhibit traits (and occupy ecospace) inaccessible to aquatic lifeforms. On the whole, there is a distinctive trend favouring the emergence of high-performance innovations on land relative to water – more specifically, 11 out of the last 13 major post-Ordovician breakthroughs appear to have originated on land (Vermeij, 2017). Of these 13 innovations, the dispersal of propagules (e.g. spores and seeds) by animals, and the communal construction of dwellings have not yet arisen (or been documented) in oceans.

The question of whether 'forbidden phenotypes' (Vermeij, 2015) in water, i.e. external characteristics that are unique to land-based organisms (on Earth), can eventually arise on ocean planets patently lacks a conclusive answer at this stage. Nevertheless, based on the available empirical evidence from Earth elucidated earlier, the majority of the higher evolutionary transitions might not occur on planets with (sub)surface oceans.[23] Although cetaceans have been (controversially) linked with certain 'human' traits such as culture, consciousness and intelligence (Griffin, 2001; Rendell and Whitehead, 2001; Whiten and van Schaik, 2007; Marino *et al.*, 2007; Whitehead and Rendell, 2015; De Waal, 2016; Whitehead, 2017) - see, however, Tyack (2001); Penn *et al.* (2008); Richerson and Boyd (2008); Manger (2013); Suddendorf (2013) for alternative perspectives – at this stage, it remains fundamentally unclear as to whether certain attributes including advanced tool construction, mental time travel, recursive thought processes, and *perhaps* syntactical–grammatical language (Roth and Dicke, 2005; Suddendorf and Corballis, 2007; Corballis, 2011; Berwick and Chomsky, 2016; Laland, 2017) may constitute evolutionary innovations that are unique to land.

[23]In this context, it is worth pointing out that most planets with $R \gtrsim 1.6R_{\oplus}$ are not likely to possess a rocky composition (Rogers, 2015; Chen and Kipping, 2017). Moreover, many exoplanets in the HZ of M-dwarfs could also end up as ocean planets (Tian and Ida, 2015).

Implications for detection and panspermia

We will briefly explore some of the implications stemming from the likelihood of life on planets with subsurface worlds, and comment on the prospects for detection.

Number of planets with potential subsurface oceans

We begin by introducing some notation. The variable 'N' denotes the number and '$\mathcal{P}$' signifies the probability. We use subscripts 'HZ' and 'SO' to distinguish between rocky planets located in the HZ and those that could have subsurface oceans, i.e. type B and U planets introduced in Sec. 2.2. We will introduce the rest of the notation as we proceed further. However, before doing so, let us recall that all planets in the HZ are not guaranteed to have water on the surface; similarly, not all type B and U planets will actually possess subsurface oceans.

In order to estimate N_{HZ}, we can use the data from the *Kepler* mission. Statistical studies have yielded fairly disparate results depending on the spectral type of star considered, the limits of the HZ, etc. A summary of these findings for main-sequence stars and white dwarfs can be found in Table 1 of Kaltenegger (2017) and Section 6 of van Sluijs and Van Eylen (2018), respectively. We adopt an estimate of ~ 0.1 rocky planets in the HZ per host star. This value is slightly on the conservative side, since it is approximately 50% of the corresponding fraction for M-dwarfs (Dressing and Charbonneau, 2015), which are the most common type of stars in our Galaxy. Thus, using the fact that there are $\sim 10^{11}$ stars in the Galaxy, we obtain

$$N_{\mathrm{HZ}} \sim 0.1 \times 10^{11} \sim 10^{10}. \tag{51}$$

Next, let us estimate N_{SO}, i.e. the number of planets that could host subsurface oceans. This can be done by noting that $N_{\mathrm{SO}} \approx N_{\mathrm{B}} + N_{\mathrm{U}}$, where N_{B} and N_{U} are the number of bound and free-floating planets, respectively. However, it is not easy to estimate how many potential planets with subsurface oceans exist outside the HZ of the host star. Hence, at this stage, we must resort to a variant of the Copernican Principle, also referred to as the Principle of Mediocrity.[24]

We will therefore assume that the solar system is not highly atypical, and that the number of potential subsurface worlds with oceans per star is similar to that of the Solar System. Since we are not concerned with the actual existence of subsurface oceans herein (merely the possibility that they could have one), we count the number of objects within the range $200 < R < 6400$ km in the solar system outside the HZ;[25] the lower and upper bounds reflect the radius of Enceladus and Earth

[24]It shares close connections with the Principle of Cosmic Modesty proposed recently (Loeb, 2017), but their stances are not exactly the same.

[25]http://www.johnstonsarchive.net/astro/tnoslist.html

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respectively. Bearing in mind the fact that not all TNOs have been detected, we find that there are ~ 100 'planets' within the above range. If the lower cutoff is increased to ~ 500 km, the number drops to ~ 25. Barring Enceladus and possibly Mimas, most of the objects that may possess subsurface oceans have $R \gtrsim 500$ km (Lunine, 2017), and therefore it seems more prudent to use the higher cutoff. With this set of assumptions, we arrive at

$$N_{\rm B} \sim 25 \times 10^{11} \sim 2.5 \times 10^{12}. \quad (52)$$

In order to determine $N_{\rm U}$, we can adopt two different strategies. The first is to use results from simulations, whereby the ejected number of planets (with different masses) is computed for a wide range of initial planetary and debris disc configurations. Here, we need to impose a lower cutoff once again for free-floating planets that could host subsurface oceans. For sufficiently high-volatile inventories, planets similar in size to Europa should be able to retain oceans purely through radiogenic heating (Spohn and Schubert, 2003). From Figure 1 of Barclay *et al.* (2017), it can be seen that ~ 100 planets above this cutoff are ejected from systems with giant planets and ~ 10 planets when there are no giant planets. Since giant planets exist only in $\sim 20\%$ of all stellar systems, on average ~ 30 planets are ejected per star, thus yielding

$$N_{\rm U} \sim 30 \times 10^{11} \sim 3 \times 10^{12}. \quad (53)$$

In the above calculation, we have assumed that there exist $\sim 10^{11}$ stars in the Milky Way. In reality, the total number of stars that have existed over the Milky Way's lifetime is higher, especially given that the star formation rate peaked at $z \approx 1.9$ (Madau and Dickinson, 2014), and type U planets would have been ejected from them.[26] However, we shall proceed with the more conservative estimate, namely (53), in our subsequent analysis.

The second method for inferring $N_{\rm U}$ relies upon the very recent discovery of the putative interstellar asteroid 'Oumuamua by the Pan-STARRS telescope (Meech *et al.*, 2017).[27] There have been several follow-up studies concerning the structure, origin and travel time of this asteroid (Mamajek, 2017; Gaidos *et al.*, 2017; Ye *et al.*, 2017; Jewitt *et al.*, 2017; Bannister *et al.*, 2017; Bolin *et al.*, 2018) and its implications for the formation and architecture of planetary systems (Trilling *et al.*, 2017; Laughlin and Batygin, 2017; Raymond *et al.*, 2017; Jackson *et al.*, 2017; Ćuk, 2018). We make use of the fact that the density of such objects has been predicted to be $\sim 10^{14} - 10^{15}$ pc^{-3} in the solar neighbourhood (Portegies Zwart *et al.*, 2017; Feng and Jones, 2018; Do *et al.*, 2018). It translates to a value of $\sim 0.01 - 0.1$ AU^{-3}, which is about 1–2 orders of magnitude higher than McGlynn and Chapman (1989), 2–3 orders of magnitude larger compared with Engelhardt *et al.* (2017) and 6–7 orders of magnitude higher than Moro-Martín *et al.* (2009). Collectively, these estimates serve to illustrate the fact that there is significant uncertainty surrounding the number of such objects in the solar neighbourhood (Cook *et al.*, 2016).

In order to carry out the order-of-magnitude calculations, we assume that the value specified above serves as the global density in the Galaxy. Upon doing so, we find a total of $\sim 7 \times 10^{25}$ objects in the Milky Way. In order to compute the number of objects with diameter >3000 km, we resort to the Copernican Principle and use Figure 1 of Bottke *et al.* (2005) to formulate an approximate power law with spectral exponent ≈ -2.5 but, in reality, there exists significant variability based on the asteroid size, age, etc. that will not be considered here. We note that our choice of the power-law distribution is comparable with the size distribution of elliptic comets, but less steep compared with Kuiper Belt Objects (Moro-Martín *et al.*, 2009). Since the diameter of 'Oumuamua is ~ 0.1 km, we find that

$$N_{\rm U} \sim 7 \times 10^{25} (3 \times 10^4)^{-2.5} \sim 4 \times 10^{14}, \quad (54)$$

which is higher than (53) by two orders of magnitude; this translates to $\sim 10^3$ type U (Moon-sized and larger) worlds per star which is approximately equal to the prediction of $\gtrsim 2 \times 10^3$ free-floating objects per star in Dai and Guerras (2018), but is about two orders of magnitude lower than the estimate provided in Figure 1 of Strigari *et al.* (2012). Nevertheless, we shall adopt the more conservative value, given by (53), in our subsequent analysis.

Thus, upon adding (52) and (53), we arrive at $N_{\rm SO} \sim 5.5 \times 10^{12}$. From (51), one can see that $N_{\rm SO}$ is $\sim 10^3$ times higher than $N_{\rm HZ}$. Hence, one can pose the question: since planets with subsurface oceans are more common than rocky planets in the HZ, why do we find ourselves on the latter? The reason most likely stems from the fact that 'we' refers to an intelligent, conscious and technologically sophisticated species. Hence, it is still plausible that the probability of life on these subsurface worlds ($\mathcal{P}_{\rm SO}$) is non-negligible, but the likelihood of technological life could instead be selectively lowered on type U and B planets as discussed in Sec. 4.3.[28]

On the likelihood of lithopanspermia

Lithopanspermia represents the transfer of life by means of rocky material from one object to another (Burchell, 2004; Wesson, 2010; Wickramasinghe, 2010; Lingam, 2016a). Most studies have tended to focus on either interstellar (Napier, 2004; Belbruno *et al.*, 2012) or interplanetary panspermia (Melosh, 1988; Gladman *et al.*, 1996, 2005).[29] Here, we will briefly explore the possibility that free-floating planets could seed life on gravitationally bound planets through panspermia (Lingam and Loeb, 2018a); a variant of this idea was also discussed in Wickramasinghe *et al.* (2012). Our subsequent discussion is also applicable, with some slight modifications, to the scenario wherein type U planets may facilitate the transfer of biomolecules by means of pseudo-panspermia (Orgel, 2004; Lingam and Loeb, 2017a).

We envision a two-step process wherein a free-floating planet is temporarily captured by a star, and then seeds other planets orbiting that star. The total probability $\mathcal{P}_{\rm tot}$ for this process is

[26]Type U planets are in sharp contrast to rocky planets in the conventional HZ, whose habitability will be terminated even prior to the star's death (Rushby *et al.*, 2013).

[27]https://www.nasa.gov/feature/jpl/small-asteroid-or-comet-visits-from-beyond-the-solar-system

[28]A related question concerning the likelihood of intelligent life on planets around M- and G-type stars was discussed in Waltham (2011); Loeb *et al.* (2016); Haqq-Misra *et al.* (2018), with some potential solutions having been advanced in Dong *et al.* (2017a,b, 2018) and Lingam and Loeb (2017d,e).

[29]Panspermia in the Galactic centre (Chen *et al.*, 2018) and in globular clusters (Di Stefano and Ray, 2016) can be viewed as a juxtaposition of the interstellar and interplanetary cases because of the close distances between the stars, although the likelihood of planet formation and stable orbits in such environments may be quite low (de Juan Ovelar *et al.*, 2012; Penny *et al.*, 2016).

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estimated through a Drake-type equation:

$$\mathcal{P}_{\rm tot} = \mathcal{P}_{\rm cap} \cdot \mathcal{P}_{\rm planet} \cdot \mathcal{P}_{\rm SO} \cdot \mathcal{P}_{\rm PS}, \tag{55}$$

where $\mathcal{P}_{\rm cap}$ is the capture probability of a free-floating planet by a star in its lifetime, $\mathcal{P}_{SO}$ is the probability that the captured planet already has life, $\mathcal{P}_{\rm planet}$ is the number of planets that could host life around that star (but not necessarily inside the HZ), and $\mathcal{P}_{\rm PS}$ is the probability of interplanetary panspermia. Note that $\mathcal{P}_{\rm planet} \sim 0.1$ if we restrict ourselves only to rocky planets inside the HZ. However, allowing for the possibility of subsurface ocean worlds, we set $\mathcal{P}_{\rm planet} \sim 1$. It is not easy to properly assess $\mathcal{P}_{\rm cap}$ since it will depend on the velocity dispersion, stellar and planetary masses, inclination angle, etc. Recent simulations undertaken by Goulinski and Ribak (2018) appear to suggest that $\mathcal{P}_{\rm cap} \sim 0.01$, but this value is parameter-dependent.

Using these values, we find $\mathcal{P}_{\rm tot} \sim 0.01 \cdot \mathcal{P}_{\rm SO} \cdot \mathcal{P}_{\rm PS}$, and the magnitudes for the remaining two variables are highly uncertain. Since $\mathcal{P}_{\rm PS}$ represents the probability of interplanetary panspermia,[30] its likelihood of occurrence should be relatively higher compared with interstellar panspermia (Melosh, 2003); for instance, Mileikowsky *et al.* (2000) found that the transfer of microbes from Mars to Earth was highly probable, and the transfer from Earth/Mars to the Galilean satellites is also possible, especially during the Late Heavy Bombardment (Worth *et al.*, 2013). Furthermore, for low-mass stars with closely packed planetary systems, the chances for interplanetary panspermia are boosted by several orders of magnitude compared with our Solar System (Lingam and Loeb, 2017a). In contrast, the prospects for interstellar panspermia appear to be much lower (Adams and Spergel, 2005). Finally, we are left with $\mathcal{P}_{\rm SO}$, and there is no available method to estimate it. If life is discovered on Europa or Enceladus, it will help constrain the probability of life originating on planets/satellites with subsurface oceans. The implications of discovering life elsewhere for the timescale associated with the origin of life have been discussed in Spiegel and Turner (2012).

In order to assess the total number of stellar systems that have been seeded with panspermia, we must multiply $\mathcal{P}_{\rm tot}$ with 10^{11}, which is a fairly large number. Hence, even if we choose $\mathcal{P}_{\rm SO} \sim 10^{-3}$ and $\mathcal{P}_{\rm PS} \sim 10^{-2}$, we find that $\sim 10^4$ stellar systems could have been seeded with life. These are, of course, fiducial estimates and the actual number of stellar systems seeded can be either much higher ($<10^9$) or much lower (perhaps equal to zero). Future statistical surveys can constrain the likelihood of panspermia processes by looking for signs of clustering; it was pointed out in Lin and Loeb (2015) – see also Lingam (2016b) – that the detection of $\gtrsim 25$ planets with biospheres will enable (under ideal circumstances) a rigorous test of the panspermia hypothesis.

The prospects for detection

We will confine ourselves to discussing the detection of type U planets herein, since the case for *in situ* exploration of type B worlds is more straightforward to espouse given the relatively large number of moons and planets with subsurface oceans in our Solar System (Nimmo and Pappalardo, 2016). A possible difference between type U worlds and type B planets/moons within our Solar System is that the initial conditions for their formation might have been different, e.g. gas-starved versus gas-rich discs. In turn, such distinctions could have important consequences for the subsequent geological, chemical and biological evolution of these worlds, thereby providing a motive for the detection and study of type U worlds.

From (53), we see that the number of free-floating planets (with $R \gtrsim 0.3 R_\oplus$) is about 30 times higher than the total number of stars. Since the nearest star is ~ 1 pc away, we suggest that the nearest such object might be located at a distance of $\langle r \rangle \sim 0.01 - 0.1$ pc from the Earth, which translates to $\langle r \rangle \sim 2 \times 10^3 - 2 \times 10^4$ AU. Note that the lower bound is roughly comparable with the inner edge of the Oort cloud and the aphelion of certain TNOs such as Sedna.

We can estimate the thermal flux from this planet that would be received on Earth using

$$S_{\rm max} \approx 1.5 {\rm mJy} \left(\frac{T_s}{40 {\rm K}} \right)^3 \left(\frac{R}{R_\oplus} \right)^2 \left(\frac{\langle r \rangle}{2000 {\rm AU}} \right)^{-2}, \tag{56}$$

where the flux density $S_{\rm max}$ has been computed at the black body peak (Wien maximum), with $\lambda_{\rm max} \approx 126\, \mu$m for the characteristic temperature of 40 K for type U planets. It can be seen that the value of $\lambda_{\rm max}$ is in the far-IR range, and several telescopes are operational at such wavelengths.

The maximum distance at which Earth-sized free-floating planets can be detected is ~ 830 AU (Abbot and Switzer, 2011) for both LSST (Jones *et al.*, 2009) and PAN-STARRS (Jewitt, 2003), suggesting that the characteristic distance of the nearest type U planet falls below the detection threshold. The maximum sensitivity of the Herschel/PACS instrument appears to be a few mJy (Berta *et al.*, 2010; Poglitsch *et al.*, 2010), indicating a borderline case. In contrast, the Cornell–Caltech Atacama Telescope (CCAT) has been predicted to reach a sensitivity of ~ 0.36 mJy for a wavelength of 200 μm,[31] which is slightly lower than the value of $S_{\rm max}$ obtained above. Hence, it seems plausible that upcoming telescopes could detect such free-floating planets although it is apparent that the result will depend critically on the number density of such objects in the solar neighbourhood.

The question as to whether any tangible biomarkers, or even the interior composition and structure (Vance *et al.*, 2018), pertaining to these planets can be unambiguously identified is not easy to resolve. The major difficulty stems from the fact that there is no atmosphere, since most studies have hitherto focused on atmospheric biosignatures such as oxygen and ozone (Meadows, 2017; Grenfell, 2017). If these planets emit plumes akin to Enceladus (Waite *et al.*, 2009; Postberg *et al.*, 2011) and perhaps Europa (Sparks *et al.*, 2016, 2017), it may be possible to search for biomarkers therein (McKay *et al.*, 2008; Judge, 2017) but a significant difficulty arises from the fact that the photon flux received at Earth scales as $\langle {\rm r} \rangle^{-2}$. Hence, *ceteris paribus*, the flux from the nearest Type U planet would be $10^{-5} - 10^{-6}$ lower than that of Enceladus unless we serendipitously discover such an object much closer to our planet. Another possibility in the future is to make use of small spacecraft powered by light-sail technology, along the lines of the recently announced *Breakthrough Starshot* project,[32] for carrying out flyby missions (Hein *et al.*, 2017). A spacecraft travelling at

[30] More specifically, it refers to the probability of glaciopanspermia, i.e. life being seeded by ejecta from impacts of icy worlds; a similar process has been conjectured to have occurred in the early Solar System, with Ceres functioning as the source (Houtkooper, 2011).

[31] http://www.ccatobservatory.org/index.cfm

[32] http://breakthroughinitiatives.org/initiative/3

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20% the speed of light might be able to reach the nearest type U planet in a span of ~ 1 yr.

We conclude by pointing out a promising, and possibly universal, feature of life-as-we-know-it: its propensity to generate thermodynamic disequilibrium (Barge *et al.*, 2017; Branscomb *et al.*, 2017). This feature has been invoked widely in the context of detecting atmospheric biosignatures (Kaltenegger, 2017), with some caveats (Krissansen-Totton *et al.*, 2016), ever since the pioneering work by Lederberg (1965) and Lovelock (1965). However, the atmosphere is only one component of the Earth that exists in a state of disequilibrium, and it has been recognized that the energy balance of the surface is also profoundly altered by a mature biosphere (Lovelock and Margulis, 1974; Schwartzman, 1999; Lenton and Watson, 2011; Kleidon, 2016).[33] Hence, it remains an open question (and one perhaps worth further consideration) – especially given our rudimentary understanding of planetary bioregulation mechanisms (Tyrrell, 2013) – as to whether subsurface exolife would be capable of altering the planetary interior to a degree that it can be detected with sufficient precision to distinguish it from false positives (Walker *et al.*, 2017).

Conclusions

The goal of this work was to examine the constraints on the habitability of planets and moons with subsurface oceans and an outer ice envelope. Some of our findings are also applicable to planets with deep terrestrial biospheres or ocean planets (with surface water). We began by presenting a simple model of a conducting ice layer, and showed that its thickness was regulated by the mass, surface temperature and the availability of radioactive materials. As a result, we concluded that a wide range of 'planets' with ice shells of moderate thickness may exist in a diverse array of habitats.

Although the availability of water is an important constraint, life-as-we-know-it also requires an energy source for both origination and sustenance. Hence, we quantified the energy available from a wide range of sources such as ionizing radiation, exogenous delivery of dust, and radiogenic heating. In each instance, we computed the amount of amino acids that could be produced. However, there are several steps between the formation of prebiotic compounds and the origin of life, and we examined how certain unique properties of ice can play a beneficial role in this regard, especially with respect to concentration and polymerization of these molecules.

Subsequently, given these energy sources we examined the biological potential of these worlds. We found that a wide variety of mechanisms are capable of supporting biospheres, such as the energy derived from the delivery of oxidants from the surface, hydrothermal vents and the radiolysis of water. Under certain circumstances, a redox balance akin to that of Earth, and perhaps Europa, may exist although its likelihood is low in general. In most cases, we concluded that the rate of biomass production was likely to be several orders of magnitude lower than on Earth. Next, we highlighted the fact that life also requires a steady long-term supply of bioessential nutrients in addition to energy sources, and consequently crucial biogeochemical cycles (e.g. the phosphorus cycle) might face challenges on these planets. We presented a brief sketch of the major evolutionary transitions on Earth, and hypothesized that some of the later (more complex) innovations have a low probability of occurring on worlds with (sub)surface oceans.

We concluded our analysis by presenting heuristic estimates for the total number of planets capable of possessing subsurface oceans that exist in our Galaxy, and found that they are perhaps $\sim 100 - 1000$ times more common than rocky planets in the HZ of stars. We briefly discussed the possibility that free-floating planets can enable lithopanspermia to occur on an interplanetary level. We explored potential avenues for detecting these planets, and found that the identification of distinctive subsurface biosignatures does not appear to be feasible with current space- and ground-based telescopes.

To summarize, life on (exo)planets with subsurface oceans is likely to face *sui generis* challenges that are not prevalent on Earth. Examples include the lack of an abundant energy source equivalent to sunlight, and the possibility that the biosphere becomes primarily oligotrophic. On the other hand, we have not been able to identify any definitive limiters that can prevent biospheres from emerging and functioning over geologic timescales. As these worlds are likely to be far more abundant than rocky planets in the HZ of stars, we suggest that more effort should focus on modeling and understanding the prospects for life in subsurface oceans. By doing so, we will be able to take one step further towards understanding whether life, especially sentient life, in the Universe is a 'cosmic imperative' (de Duve, 1995; Morowitz and Smith, 2007; Bains and Schulze-Makuch, 2016) or a genuinely rare occurrence (Simpson, 1964; Mayr, 1985; Ward and Brownlee, 2000; Morris, 2003).

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[33]In addition to the energy balance, many crucial components of Earth's biosphere, such as productivity, nutrient cycles and even the evolutionary process itself, have been significantly influenced by the emergence of organisms (Lewontin, 2000; Odling-Smee *et al.*, 2003; Post and Palkovacs, 2009; Butterfield, 2011; Laland *et al.*, 2015).

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