The Most Common Habitable Planets II - Salty Oceans in Low Mass Habitable Planets and Global Climate Evolution

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ABSTRACT
The available models of global climate evolution in habitable earthlike planets do not consider the effect of salt content in oceans, which affects water evaporation. Two distinct categories of such planets are considered in this work: planets with deep oceans, but with intrinsically high salinities due to the weaker salt removal process by hydrothermal vents; and planets with shallow oceans, where the increase in salt content and decrease in ocean area during the onset of glaciation cause a measurable negative feedback on perturbations, helping delay the onset of ice ages. We developed a toy climate model of a habitable planet on the verge of an ice age, using a range of initial salt concentrations. For planets with deep oceans and high salinity we find a considerable decrease in land ice sheet growth rate, up to ~ 23% considering the maximum salinity range. For planets with shallow oceans, the effect of intrinsic high salinity previously modelled is reinforced by the negative feedback, to the point of effectively terminating the land ice sheet growth rate during the time-scale of the simulation. We also investigate the application of this model to the putative ocean of early Mars, and find that the results lie in between the two categories. We conclude that this new phenomenon, which can be viewed as an abiotic self-regulation process against ice ages, should be taken into account in studies of habitable planets smaller and drier than the Earth, which may well represent the bulk of habitable planets.

Key words: astrobiology – planets and satellites: oceans – planets and satellites: terrestrial planets – planets and satellites: surfaces – planets and satellites: composition

1 INTRODUCTION
Climate evolution studies of habitable planets have undergone two different waves of stimulus over the past century: the first one was in the wake of the exploration of Venus and Mars by robotic probes, concretely starting the field of comparative planetology in the 1960s (e. g., Kiefer et al. 1992; Barrie 2008; Launius 2012); the second one started with the discovery of the first extrasolar planet around a main sequence star (Mayor & Queloz, 1995), almost a quarter of a century ago. And, last but not least, the field has also received recent and strong input from studies of the climate of our own planet, with important and practical results, like long term alterations in climate patterns due to the increasing anthropogenic greenhouse effect. Nowadays the number of confirmed discovered extrasolar planets has surpassed 4200 (Schneider et al. 2011), and, with the aid of fast computing, we are able to simulate the climate of not just hypothetical, but also actual planets, constrained by data on radiation flux from the star, size and density of the planet, and, in some cases, data on atmospheric composition (Wallack et al. 2019; Molaverdikhani et al. 2020). However, most of the data available on atmospheric composition still refer to uninhabitable planets like Hot Jupiters, but observatories like the James Webb Space Telescope, the E-ELT and TESS may change this scenario in the near future, enabling the analysis of the atmospheres of as yet undiscovered nearby rocky planets in the habitable zone of their stars – a tantalizing possibility further fuelled by the discovery of a possibly habitable rocky planet in the habitable zone of Proxima Centauri, our closest neighbor in the Galaxy (Anglada-Escudé et al. 2016; Ribas et al. 2016; Turbet et al. 2016). Unfortunately, the measurement of key variables such as the water content of such planets remain beyond the current technological capabilities, and have to be assumed in simulations. The presence of water in the liquid state at planetary surfaces has been elected as a prime criterion to define their habitability, helping to frame the so called orthodox stellar habitable zone around...
main sequence stars of different masses (Kasting et al. 1993; Franck et al. 2000; Kopparapu et al. 2013). The origin of the Earth’s water and its total content are still under debate, as well as the possibility that planets similar to our own in the habitable zones of other stars may be as rich, or even richer in water than the Earth (Noack et al. 2016). On the other hand, details about water delivery to planets inside the habitable zone as well as volatile removal by high-energy emissions may play a crucial role on planetary water content. Particularly for very low mass stars, total or near total removal of planetary volatiles by prolonged exposure to XUV radiation and particle winds may result in the average habitable planet, at least at the low end of the mass spectra, for low mass bodies, being considerably water poor as compared to the Earth (Lammer et al. 2009; Ribas et al. 2016; Bolmont et al. 2017; Dong et al. 2017; Airapetian et al. 2017), even though models suggest that some degree of water retention, contingent upon initial water inventory, remains a clear possibility (Bolmont et al. 2017). Suggestions that most of the potentially habitable rocky planets are probably small as well as water deficient (Lissauer 2007; Pinotti 2013) strengthen the importance of detailed studies of such worlds, including climate models and conditions under which they might hold on to their volatile inventories and remain habitable.

If water deficient habitable planets are as numerous as these studies suggest, then the climate models would benefit by taking into account the total superficial water volume as a parameter, as well as its distribution along the planet’s area. We will see later that this distinction is quite relevant.

Moreover, smaller planets would be expected to have oceans with higher salt content, since the main salt removal process, namely the ocean ridge hydrothermal circulation system (Langmuir & Broecker 2012), would be less intense, along with tectonic activity in general, due to the faster cooling of such planets. This reasoning, coupled to the fact that water evaporation and freezing are affected by salt content, leads to the possibility that salt content may play an important role in the climate evolution of such planets. More specifically, the decrease of water evaporation would affect the formation of glacers on land, since the balance of water content in the atmosphere depends on it, and the depression of freezing point would delay the formation of sea ice. Moreover, for planets with shallow oceans, the surface seawater volume is smaller still as compared to planets with oceans with average water depth on the order of kilometers, and thus the decline of ocean volume during glaciation might be sufficient to increase salinity over time, creating a negative-feedback effect on glaciation.

In order to study the effects of ocean salinity on the climate of habitable planets with varying degrees of surface water extent and volume, we developed a toy climate model, with 26 variables and 26 algebraic-differential equations, and investigated the salinity-evaporation relation for a specific climate dynamics, namely, a planet on the verge of an ice age.

The paper is organized as follows: in section 2 we discuss the possible “pristine” surface water content of exoplanets; section 3 contains a discussion about the loss of volatiles caused by stellar wind and radiation; in section 4 we discuss the variables that determine the salt content on the oceans of the Earth and of exoplanets; in section 5 there is a description of our climate toy model; section 6 shows the results of the simulations and a discussion about them, including a customized simulation for the planet Mars; finally, section 7 presents our conclusions.

2 SURFACE WATER VOLUME IN ROCKY PLANETS

Since we will perform simulations of the climate of planets of different sizes, an a priori step is an assessment of their possible surface water volume. The problem of the origin of Earth’s oceans is still unresolved, and researchers have arguments for both wet-endogenous and exogenous sources (van Dishoeck et al. 2014; D’Angelo et al. 2019; Ida, Yamamura & Okuzumi 2019). The wet-endogenous theory links the water to solids that formed the Earth, and outgassed to the surface (Elkins-Tanton 2012; Lebrun et al. 2013; Noack et al. 2016). The exogenous one calls for water delivery by small bodies or by nebular gas (Genda and Ikoma 2008; Morbidelli et al. 2012; Raymond et al. 2014), coming from outer regions of the solar system, after the formation of the Earth. Constraints like the protoplanetary disk temperature in the region where Earth is believed to have been formed, and the deuterium to hydrogen ratio of the water in the Earth preclude a simple answer, and a combination of both sources is a possibility.

The amount of water delivered to rocky extrasolar planets by comets and asteroids is probably a stochastic process, which depends not only on the protoplanetary disk composition, but also on the number, size and radial distribution of other planets in the systems, variables with strong stochastic nature, given the available sample of extrasolar multiplanet systems and planet formation simulations.

The wet-endogenous hypothesis, on the other hand, allows simple deterministic correlations to be developed. Let’s consider the total surface water volume in a given planet by Eq. (1):

\[ V_s = h\theta\pi R_p^2 \]  

(1)

where \( V_s \) is the total volume, \( h \) is the average water depth of the oceans, \( \theta \) is the initial fraction of the planetary surface area which is covered by water, and \( R_p \) is the planetary radius. Then, the ratio between the surface water volume and the total surface area of the planet is given by \( \frac{V_s}{A_p} = h\theta \), where \( A_p = 4\pi R_p^2 \) stands for the total planetary surface area.

In our case it is important not to use one term to express the ratio, since the evaporation process depends on the water area exposed to the atmosphere, indicated by \( \theta \).

Considering also our premise that the main source of water comes from the interior of the planet, we can write a relation between this volume and the planet’s mass given by Eq. (2):

\[ V_s = \delta \frac{\rho_w}{\rho_h} \frac{\pi R_p^3}{\theta} \]  

(2)
where $\rho_p$ is the average planetary density, $\rho_w$ is the water density, and $\delta$ is the fraction of water mass compared to the total mass of the planet. Then, combining Eq. (1) with Eq. (2), we obtain Eq. (3):

$$h\theta = \delta \rho_w \frac{2}{3} \pi R_p^3$$  (3)

This equation indicates a linear relation between planetary radius and the product $h\theta$, considering $\delta$ and $\rho_p$ as relatively constant for rocky planets in the Habitable Zone. However, we must bear in mind that many factors may turn this assumption wrong, due to different planetary composition, including percentage of radionuclides, which affect volcanism and plate tectonics, and consequently outgassing efficiency. In addition, Eq (3) is probably an upper limit of water coverage, since water loss processes, mainly due to stellar activity, have an increasing importance for smaller planets.

According to Eq (3), for sufficiently high values of $R$, $\theta_0$ tends to 1, and we will have ocean planets, as described in theoretical studies (Seager & Elkins-Tanton 2008; Adams, Kaltenegger & Sasselov 2013; Kitzmann et al. 2015; Simpson 2019). However, in our work we will focus on low mass habitable planets, which tend to have both less extensive and shallower oceans.

Although the endogenous hypothesis leads to simple correlations between planet size and water coverage, its unknowns ($\delta$, $\rho_p$) restrict severely our ability to make reliable predictions. The exogenous hypothesis, while stochastic in nature, offers the possibility of a wider range of surface water content on exoplanets, and, at least currently, is considered to be the most probable mechanism that allowed the existence of liquid water on the surface of the Earth and possibly on exoplanets in general (Raymond & Ixidoro 2017). Our toy model does not depend on the origin of the surface water in exoplanets, its only premise being that planets smaller than the Earth tend to be drier, due mainly to the considerations described in section 3. Our assumptions can thus be accommodated to a varying range of sources for a planet’s water inventory. Finally, it is worth mentioning that even the amount of exogenous sources of water is expected to depend on the size of the planet, given the Hill sphere.

**3 PARENT STAR AND LOSS OF VOLATILES**

The development of the previous section gives us a direction on how dry a rocky planet may become as a function of its size. Using the Earth as the basis for extrapolation, with around 70% of its surface covered by water, and considering its average ocean depth of 3.7 km a constant, we derive, for planets with radii 75% and 50% that of the Earth’s, a water coverage of 52.5% and 35% respectively.

However, this is but a rough estimate of the initial water coverage when the planet is very young. Stellar winds, X-ray and UV (XUV) radiation tend to erode the atmosphere and volatiles from the surface of rocky planets. This process is dependent on many factors involving both planetary and stellar properties, and a large variety of end results concerning volatile retention is expected. For solar-type stars in general, with masses between 0.6 and 1.5 solar masses, the high-activity phase is very short-lived and directly connected to the evolution of rotation (Wood et al. 2002, 2005; Ribas et al. 2005; Ribas et al. 2010; do Nascimento et al. 2016) for both high-energy XUV photons and winds, and only very low mass planets are affected, as is the case with Mars. For lower mass stars inside the red dwarf regime, a much longer phase is maintained (West et al. 2008) starting around type M3. Connected to the much closer-in habitable zones for such underluminous stars, this fact makes for much higher volatile losses expected for rocky planets inside the habitable zones of stars less massive than about ~0.5 solar masses. Even more severe conditions are expected for the less massive rocky planets (our main focus in this work), since their lower gravity facilitates atmosphere loss and atmospheric replenishment from volcanism is expected to be shorter lived.

Yet another factor hinder the retention of volatiles for undermassive planets: the core dynamo which generates a protective magnetosphere is expected to be weaker from the start and subside faster than the ones of bigger planets. Indeed, Zuluaga et al. 2013 modelled the evolution of planetary magnetospheric properties for a range of assumptions about mass, chemical composition and thermal evolution, and calculated the magnetic standoff radius, which is an underestimate of the actual size of the dayside magnetosphere. They find that even planets with ~0.5 Earth masses are able to achieve and maintain sizable standoff magnetospheric radii, not much unlike planets as massive as the Earth, as long as they remain tidally unlocked. These results hold even during the star’s initial phase of strong winds. For tidally locked planets, much diminished standoff radii are found, yet still some measure of protection is to be found. Tidally locked rotation, or capture into low order rotational resonances, for planets inside the habitable zone, is essentially ubiquitous for stars less massive than about 0.7 solar masses (Porto de Mello et al. 2006), and therefore considerably larger volatile loss from various physical mechanisms are expected for rocky planets orbiting stars less massive than about 0.6-0.7 solar masses, increasing in severity as we consider less massive planets.

Considering these complications, and recognizing that actual planets probably span a much wider range of properties, we finally assume that the Earth has suffered minor volatile loss, and we opted to perform simulations for three different planetary radii relative to Earth’s (100%, 75% and 50%). Accordingly, the values of $\theta_0$ were chosen as 70%, 40% and 10% respectively, which incorporates a non-linear reduction factor on the initial values estimated form Eq. (3). Our assumed values obey a range between so-called aqua planets and land planets (Kodama et al. 2019), a distinction with some importance for the definition of the inner edge of Habitale Zones.

**4 SOURCES AND SINKS OF OCEAN SALT**

We shall consider in our model that, for lower planetary radius, plate tectonics will probably play a shorter and less intense role in many aspects of planetary climate, including its role as a sink of oceanic salt, through hydrothermal circulation. This reasoning, added to the probable smaller total water volume indicated by Eq (3), leads to the conclusion that the ocean of these worlds will probably be richer in salt, for the input of salt into the oceans through rivers is linked
to the hydrologic cycle, which in turn is a process relatively independent from plate tectonics in the time-scales considered in this work (10^2 - 10^5 years).

The above arguments lead us to expect that not only would the volume of surface water change with planetary radius, but also its composition. It was only recently, with the discovery of the ocean ridge hydrothermal circulation (Langmuir and Broecker 2012; Hovland, Rueslåtten & Johnsen 2018) that the enigma of seawater composition was solved. The continuous input of salt from rivers to the oceans must be balanced by a sink, otherwise the salt content of the oceans would have reached the saturation level in a geologically short time. Table 1 shows the compositions of some elements in rain, rivers, seawater and hydrothermal fluid. The hydrothermal systems, which remove sodium and chlorine, among other elements, and is a source of others, process the entire volume of the oceans in tens of millions of years, and is directly related to the existence of plate tectonics. In planets smaller than the Earth, plate tectonics is expected to be less intense – even absent in some cases – after some Gyr, compared to that of the Earth, leading to the conclusion that the ocean ridge hydrothermal circulation would probably be less efficient in removing salt from the oceans. The oceans, in turn, would still be receiving the flow of rivers due to the hydrological cycle, so that their salt content would be higher than the Earth’s oceans.

| Element | Rain | Rivers | Seawater | Hydrothermal Fluid |
|---------|------|--------|----------|-------------------|
| Ca      | 0.65 | 13.3   | 412      | 1200              |
| Mg      | 0.14 | 3.1    | 1290     | 0                 |
| Na      | 0.56 | 5.3    | 10770    | -                 |
| K       | 0.11 | 1.5    | 380      | 975               |
| Sulphate| 2.2  | 8.9    | 2688     | 28                |
| Cl      | 0.57 | 6.0    | 19000    | -                 |
| Si      | 0.3  | 4.5    | 2        | 504               |
| Fe      | 0.03 | 0.02   | 0.16     | 168               |
| Mg      | 0.007| 0.0002 | 41       |                   |

Table 1 – Compositions of Earth’s water (Langmuir and Broecker 2012) – concentrations in parts per million

Due to many other factors influencing plate tectonics, it is not possible to directly correlate planetary size with salt content in the oceans, so we will assume, in our simulations, a range of salt content for each planetary radius considered, between the value of Earth’s oceans (35000 ppm) and the value of near saturation (260000 ppm).

5 THE MODEL

In order to assess the influence of ocean salinity on the evaporation rate and ice formation in oceans and on land, we have developed a toy model for a planet with roughly the same characteristics of the Earth. The model performs a dynamic water mass balance between oceans, the atmosphere and land ice, starting with a state that goes in the direction of an ice age, that is, one in which ice formation rate on land is positive and unbounded.

Since the majority of rainwater on Earth falls on the oceans or is returned to the oceans through rivers, the dynamic water budget of the atmosphere will consider only the evaporation from oceans, multiplied by a factor that corrects for water not returned to the oceans, and the formation of glaciers on land. The average cloud coverage of the planet is considered to be constant, since its modeling is known as quite a difficult task, and a complete model is beyond the objective of this study. Moreover, the cloud coverage is not supposed to vary significantly during the time scale of the simulations.

The global surface energy balance will be given by Eq. (4), which states that the radiative energy flux emitted by the entire surface equals the absorbed stellar flux:

\[(1 - A)F\pi R^2 = 4\pi\sigma R^4T^4\]  \hspace{1cm} (4)

The average planetary albedo, A, will be a function of time in the simulations, as well as the area covered by ice, sea and land, each of which having a specific albedo. T is the average temperature of the upper troposphere, whose heat exchange mechanism is mainly radiative, F is the stellar bolometric radiative flux, \(\sigma\) is the Stefan-Boltzmann constant, \(\varepsilon\) is the emissivity, which is close to 0.9 in the infrared, and R is the radius of the planet.

Eq. (4) assumes that the term dA/dt, due to the accumulation of thermal energy in the atmosphere (Pinotti 2013), is negligible, that is, the variation of A over time is small enough, so that thermal equilibrium in the atmosphere is always attained and the variation of T over time is given by an algebraic relation with A. And, in fact, the time required for the ice/sea/land coverage to change appreciably is large compared with the time step for the integration of the ensemble of differential equations. Eq. (4) also assumes that the planet is a fast rotator, so that there is no significant temperature difference between the nightside and the dayside.

Similarly to the Earth, our model planet will have an inclination of the rotation axis relative to the ecliptic, and we will divide it into three distinct climatic regions: tropical, temperate and arctic. Let the inclination be the same also, 23.44°, or 0.1302π rad. If we also assume that the average stellar flux intercepted by the planet area is the one dictated by an equinox configuration (seasonal changes are not our focus), then we can calculate the values of intercepted radiation flux for each region. The results are shown in Table 2.

Note that the three regions encompass the northern portion and the southern one, for both are identical in terms of radiative balance in the equinox configuration, and consequently are considered as one in the simulations. Therefore, when a region is mentioned hereafter, we mean both the northern and southern ones added together.

| Region   | Area intercepting stellar radiation | Total area | Fraction of Total area |
|----------|------------------------------------|------------|-----------------------|
| Tropical | 1.5478R²                          | 4.9980R²   | 0.3977                |
| Temperate| 1.5055R²                          | 6.5318R²   | 0.5198                |
| Arctic   | 0.0883R²                          | 1.0367R²   | 0.0825                |
| Total for the planet | nR²                                  | 4R²          | 1                     |

Table 2 – Areas for the climatic regions of a planet of radius R, with inclination of rotation axis of 23.436° relative to the ecliptic, and at the equinox point of its orbit.
Eq. (4) will be used in a modified form for the estimation of the temperature of the upper troposphere of the three regions (\(T_a\), \(T_T\) and \(T_T\) for the arctic, temperate and tropical zones, respectively), that is, the intercepting and total areas will not be \(\pi R^2\) and \(4\pi R^2\) respectively, but the values for each region given in Table 1.

In order to calculate \(T_a\), \(T_T\) and \(T_T\), using the data on Table 1, we also need values for the albedos of the three regions. We will assume the following initial conditions:

- Each region will have its own ocean, with albedo given by Table 3, and the same initial water coverage fraction (\(\theta_o\)) which defines the simulation case; that is, the main simulation cases are defined by \(\theta_o=70\%\), \(\theta_o=40\%\) and \(\theta_o=10\%\); this water coverage area will be a variable in the simulations (\(\theta_o=\theta_o(t)\)), with \(i=1,2,3\), since the evaporation will affect the ocean volumes, and with different intensities;
- The initial average depth of the oceans will be 3700 m, similar to that of the Earth’s;
- The bare land, with albedo given by Table 3, will initially cover the remaining fraction (1- \(\theta_o\)); this coverage will also change over time, as the seas shrink due to evaporation;
- The tropical region will be free of sea or land ice in the model, since its surface temperature will always be above the freezing point of water;
- The temperate region will have an initial ice coverage of 1% over land (fractio of total temperate area=0.01*(1- \(\theta_i\)) and 1% over sea (fractio of total temperate area =0.01*(\(\theta_i\))); both ice coverages will be time variables in the model;
- The arctic region will have an initial ice coverage of 15% over land (fractio of total arctic area=0.15*(1- \(\theta_i\)) and 15% over sea (fractio of total arctic area =0.15*(\(\theta_i\))); both ice coverages will be time variables in the model; these values are compatible with arctic region with average surface temperature well below 0°C, as will be shown below;
- The whole planet will have a fixed cloud coverage of 30%, evenly distributed over sea and land; this value is below the present day Earth, because we will be dealing with a planet entering an ice age, with a drier atmosphere.

| Surface constitution | Albedo |
|----------------------|--------|
| Sea water            | 0.06   |
| Land                 | 0.27   |
| Ice                  | 0.35   |
| Cloud                | 0.5    |

Table 3 – Albedos for each surface constitution

With these assumptions, and applying Eq. (4) separately to each climatic zone, we are able to calculate the initial albedos and upper troposphere temperatures for each region, and for each value of initial \(\theta_o\), which defines each simulation case. The results are shown in Table 4 and Table 5.

The significant temperature differences between the regions is due to the assumption that they are independent of each other. On Earth, the differences are dampened by air and ocean currents. But there will be no sea currents in our model, since the seas are isolated; air currents are indirectly assumed (This topic will be discussed below), but the air has less heat capacity than water, so the model of three isolated regions is probably not far from the reality of a planet with the characteristics described in this work.

| Region   | \(\theta_o=0.7\) | \(\theta_o=0.4\) | \(\theta_o=0.1\) |
|----------|------------------|------------------|------------------|
| Tropical | 0.2361           | 0.2802           | 0.3243           |
| Temperate| 0.2377           | 0.2813           | 0.3250           |
| Arctic   | 0.2599           | 0.2974           | 0.3349           |

Table 4 – Initial albedos for each region and each value of \(\theta_o\)

| Region   | \(\theta_o=0.4\) | \(\theta_o=0.1\) |
|----------|------------------|------------------|
| Tropical | 281.9            | 273.4            |
| Temperate| 261.7            | 253.8            |
| Arctic   | 202.5            | 197.2            |

Table 5 – Initial upper troposphere temperatures (K) for each region and each \(\theta_o\)

The next step is the estimation of the initial temperature of the lower troposphere, which is different than that of the upper troposphere due to the greenhouse effect and air currents. We will assume that the increment, due to a greenhouse effect, is given by \(\psi\). The greenhouse effect is a function of the amount of greenhouse gases in the atmosphere. Since \(\text{H}_2\text{O}\) is a strong greenhouse gas, we will link \(\psi\) to the relative humidity of the atmosphere, \(\phi\) (\(\psi=\psi(\phi)\)), which will also be a time variable (\(\phi=\phi(t)\)). We will consider that other greenhouse gases will not change appreciably during the time span of the simulations, so that \(\psi=c+f(\phi)\). The term \(f(\phi)\) will be linear with respect to \(\phi\), that is, \(\psi=c+b\phi\), where \(b\) and \(c\) are constants. This is an admittedly simplistic proxy to the more complex relation between relative humidity and the greenhouse effect, but, then again, our toy model does not presume to be a rigorous one as a whole, and uses simplifications of phenomena and reduction of dimensions, in order to probe a specific relation, that is, the effect of ocean salinity on evaporation (this one modeled with rigorous equations), and, as a consequence, on glaciation. We did not include variation of \(\text{CO}_2\) content, for its dynamics during the onset of ice ages on the Earth is similar to that of temperature and humidity, and its greenhouse effect can be viewed as accommodated in our linear relationship with humidity. Moreover, intrinsic \(\text{CO}_2\) variability has time-scales higher than the ones studied in this work.

We will assume that \(\phi\) will be the same for all the three regions, that is, we are implicitly assuming that the air currents will transport humidity through the regions and level off any difference. This simplification carries a precision price with it, but since the objective of the toy model is to probe the correlation between the evaporation of salt water and the formation of ice sheets on land, and that the main water reservoir lies in the oceans, rather than in the atmosphere, we believe that the results will not be affected significantly.

Finally, the value of \(c\) will be different for each region, thus accounting for air currents, so that the difference between regions becomes less steep and more realistic. Table 6 summarizes the initial values for the temperature of the lower troposphere, for each region. The temperatures of the lower troposphere will be considered as the surface temperatures of the planet.

The tropical and temperate regions will have initial surface temperatures above 0°C, and will provide the atmosphere with water vapor through
ocean evaporation, whereas the arctic region will have, on the other hand, temperature well below 0 °C, working essentially as a cold trap.

| Region    | \( \theta_0 = 0.7 \) | \( \theta_0 = 0.4 \) | \( \theta_0 = 0.1 \) |
|-----------|------------------|------------------|------------------|
| Arctic    | 242.5            | 239.9            | 237.2            |
| Temperate | 291.7            | 287.9            | 283.8            |
| Tropical  | 308.9            | 304.7            | 300.4            |

Table 6 – Initial surface temperatures (K) for each region, for each value of \( \theta_0 \).

The next step, once the initial conditions for the simulations are set, is to model the dynamic water exchange between oceans, the atmosphere and land ice sheets that advance without constraints in our scenario. We consider that the source of formation of sea ice is the condensation of sea water only.

The evaporation rate is known to be directly proportional to the difference between the water vapor pressure of the air in contact with water and the water vapor pressure of the air at a given height (Harbeck 1955; Perry 1997; Babkin 2019). If we also assume that the air in contact with the liquid is saturated, then we can write the following equation for the evaporation rate \( Q \):

\[
Q = \alpha (p^* - p) \tag{5}
\]

where \( \alpha \) is a constant, \( p^* \) is the water vapor saturation pressure, and \( p \) is the water vapor pressure, both at the surface temperature. The water vapor saturation pressure is a function of the temperature only, and we will use the well established Buck equation (Buck 1981):

\[
p^* = \exp \left( \frac{18.678}{234.5} - \frac{T}{(237.144 + T)} \right) \tag{5}
\]

where \( p^* \) given in kPa, and \( T \) in degrees Celsius. On the other hand, \( p \) is related to the relative humidity by \( \phi = p/p^* \), so that Eq. (5) can be rewritten as

\[
Q = \alpha \phi (1 - \phi) \ldots \ldots \ldots \ldots (6)
\]

In order to estimate the value of \( \alpha \), we will assume an evaporation rate of the order of \( 10^3 \) mm year\(^{-1}\) for sea water at 20 °C, a value compatible with the average evaporation rate of the oceans of the Earth (Babkin 2019). If we also assume an initial average relative humidity for the toy model as 60% (\( \phi_0 = 0.6 \)), then \( \alpha = 1300 \) mm year\(^{-1}\) kPa\(^{-1}\). We will discuss later the effect of the choice \( \alpha \) on the results of the simulations.

Here we can appreciate the possible negative feedback on climate by the sea salinity, for the water vapor saturation pressure decrease with an increase in salinity, affecting directly the evaporation rate, and limiting the vapor supply (in fact the only supply) to the atmosphere. As the seas evaporate, the salinity of the water increases, and the evaporation rate decreases (for a fixed value of \( \phi \)). Over time, as more vapor is fixed in the cold traps of the polar regions and in the land and sea ices of the temperate regions, \( \phi \) will tend to get lower, and the evaporation rate will tend to get higher again, but the changes in albedo (land and land ice will advance over sea - the ice-albedo positive feedback), will force surface temperatures down, which will also affect water vapor saturation pressure. The simulations will indicate the net effect of this negative feedback over time.

Next, we need a quantitative relation between the salinity of water and the depletion of the water vapor saturation pressure \( p^* \). We assumed a linear relation between the logarithm of salinity and the logarithm of the relative vapor pressure depletion, using as data the depletion in sea water and in the waters of the Great Salt Lake; the resulting corrected (depleted) water vapor saturation pressure is

\[
p' = p^* [1 - 10(\log(\text{salinity}) - \beta/T)] \tag{7}
\]

where salinity is given in parts per million in weight (ppm), \( \beta = 5.980 \) and \( \gamma = 0.831 \). This equation agrees very well with established correlations used in the industry (Sharqawy, Lienhard V & Zubair 2010; Nayar et al. 2016).

The total water evaporation rate from the oceans, in volume per time, will be:

\[
Q = \sum [K p^* (1 - \phi) A_i] \tag{8}
\]

where \( i \) represents the climatic region (\( i = 1, 2, 3 \)) and \( A \) is the surface area of each of the three oceans (\( A_i = A_i (\theta_0(t)) \)). The values of \( p \) will evolve with the surface temperature \( T_{surf} (p' = p^* (T_{surf}(t))) \). At this point it is interesting to note that there is a positive feedback involved with the vapor pressure depletion with increased salinity and the consequent decrease in evaporation rate. For by limiting the vapor supply to the atmosphere, the greenhouse effect caused by water vapor will tend to be lower (\( \phi \) will be lower), lowering in its turn the surface temperatures. Also, as the oceans lose water and recede, the area occupied by them will diminish, exposing more bare land, which has a higher albedo, and consequently lowering the surface temperature by means of the radiative balance discussed above. The relation between the volume of each ocean and its evaporation rate is straightforward:

\[
\frac{dV_i}{dt} = -K p^* (1 - \phi) A_i \tag{9}
\]

However, the dynamic relation between the surface area of the ocean and its volume is an incognita, since we do not know the particulars of the topography of the ocean bed of our putative exoplanets, in spite of the fact that theoretical work has been done to model planetary topography (Landais, Schmidt & Lovejoy 2019). The initial ocean surface area and its volume has been calculated based on an average ocean depth of \( h_0 = 3.7 \) km, similar to that of the Earth, and the value of \( \theta_0 \). Considering that shallow areas become exposed more rapidly than the deeper ones as the oceans evaporate, we adopted a generic relation between the surface area and the volume using an exponential, that is,

\[
A_i = A_{i0} \exp \left[ -c (V_i/V_{i0}(t)) \right] \tag{10}
\]

For small volume variations, which is the case of the scenarios of deep oceans, the exponential term can be approximated as a linear term with \( V_i/V_{i0}(t) \).
The water mass balance between the oceans, the ice glaciers on land and the atmosphere will define the evolution of the relative humidity \( \phi(t) \), that is,

\[
\frac{dM}{dt} = [\rho I Z \dot{Q}(t) - \rho J \dot{G}(t)] \tag{11}
\]

where \( G(t) \) is the land ice area formation rate, \( \rho I \) is the density of water, \( \rho J \) is the density of ice, \( I \) is the average height of land glaciers, and \( J \) is the fraction of evaporated water that does not return to the oceans in the form of rain, and rivers. On Earth the value of \( z \) is around 0.004 (Henshaw, Charlson & Burges 2000), and we chose the value of 0.0035 for our simulations. The term \( M(t) \) represents the water mass in the atmosphere, which depends on the volume of the atmosphere and its temperature, which itself is a function of time (through the radiative balance) and defines the saturation vapor pressure, which in turn is related to the relative humidity by \( \phi = p/p^* \). With the value of \( p \), the volume of the atmosphere and the ideal gas law (\( pV = nRT \)), we calculate the total water mass in the atmosphere, and its derivative, which can be rewritten as \( dM/dt = f(\phi(t)) \).

The atmosphere we consider will be the lower troposphere, with its temperature dictated by the radiative balance of the upper troposphere plus the greenhouse effect. Since all climatic regions will have its own distinct lower troposphere temperature, but the same humidity, the water vapor mass in the atmosphere will be the sum of the mass of the three regions. As for the height of the lower troposphere, necessary for the calculation of the volume, we estimated its value by assuming that, with the initial simulation conditions for \( \theta_0 = 0.7 \), the total water vapor in the lower troposphere coincides with the water vapor in the atmosphere of the Earth (\( 1.3 \times 10^{20} \text{ km}^3 \), see Henshaw, Charlson & Burges 2000); the result is 1.67 km. For the smaller planets with \( R = 0.75R_{\text{Earth}} \) and \( R = 0.5R_{\text{Earth}} \) we applied a non-linear reduction factor of 0.1 and 0.01 respectively, in order to account for the erosion of the atmosphere by stellar activity, and bearing in mind that Mars’ atmosphere is around 99% thinner than the Earth’s at the surface of the planet.

Planets of different sizes with the same initial conditions should in principle present the same dynamic behavior of intensive variables (salinity, ice and water coverage as fraction of total planet area), since the relative areas are independent of planet size. The non-linear variation of atmospheric thickness (and its associated greenhouse effect) with radius would in principle disrupt this similarity. However, our simulations indicated that the results, when altering only the atmospheric thickness, are practically the same, and that is because the water mass holdup in the atmosphere is very small compared to the fluxes involved. For example, on Earth the turnover time for water in the atmosphere is only 8 days. Therefore, the results of the simulations, using as the main parameter \( \theta_0 \), can be interpreted as valid for planets of different sizes. This is also why we used three values of salinity for each \( \theta_0 \), since there probably are exceptions to our assumption that smaller means drier, with small planets with a higher ocean coverage, which developed high ocean salinities due to the lack or weakness of plate tectonics and its associated hydrothermal systems, may well exist.

The land ice formation rate was defined as being proportional to the difference between the surface temperature and the freezing point (273.15 K). Also, the availability of water vapor in the atmosphere for ice formation must be considered, but since the particulars of weather are beyond the scope of this work, we used an exponential law, so that \( G(t) \) was modeled as:

\[
G(t) = Y(273.15 - T) \exp(-x/\phi(t)) \tag{12}
\]

where \( Y \) and \( x \) are constants, with values compatible with growth rate similar to the ones observed on the onset of ice ages on Earth (10^4 years), and on planets with \( \theta_0 = 0.7 \).

The sea ice evolution does not alter the water budget calculated in the simulations, since its thickness is small and it is formed mainly from sea water. However, sea ice area does influence the albedo of the arctic regions, and consequently the surface temperature through the radiative balance. The sea ice formation rate used in the simulations is proportional to the difference between the surface temperature and the freezing point of water:

\[
\frac{dM_{\text{ice}}}{dt} = \Delta (T_f - T_s) \tag{13}
\]

where \( \Delta \) is a constant, \( T_f \) is the freezing point and \( T_s \) is the surface temperature. The freezing point of water suffers a depression due to the colligative effect by dissolved salt, according to

\[
T_f = 0 - K_f b t \quad \text{(in °C)} \tag{14}
\]

where \( K_f \) is the cryostatic constant (for water the value is 1.853 °C kg/mol), \( b \) is the molality of the solution, and \( t \) is the van’t Hoff factor. For NaCl the factor is 2. Hence, for more saline oceans, the rate of sea ice formation will decrease due to the decrease of \( T_f \).

The resulting ensemble of 26 algebraic-differential equations and variables was translated to Matlab and integrated by the DASSL package (Secchi 2010). The simulations are stopped when the sea and land in the arctic regions become fully covered with sea ice and land ice sheets respectively.

6 RESULTS

The results of the simulations are shown separately for the cases of deep oceans (initial average depth of 3.7 km - section 6.1), shallow oceans (initial average depth of 0.5 km - section 6.2) and for a hypothesized ancient martian ocean (section 6.3).

6.1 Deep oceans

Figures 1 and 2 show the results for the most Earth-like scenario of \( \theta_0 = 0.7 \). Figure 1 shows the evolution of the fraction of arctic ocean area covered by ice, for the three salinities selected, and Figure 2 shows the evolution of the fraction of arctic land area covered by ice sheets, for the same set of salinities.
The effect of freezing point depression, due to increased water salinity, on the necessary time to freeze completely the arctic oceans is evident; for a salinity of 260000 ppm the time-scale almost doubles, compared with a planet with an ocean salinity of 35000 ppm. In the case of the advancing ice sheets on land, the effect of the oceans’ salinity is not as spectacular, but there is a significant increase of 23% in the necessary time for them to cover the entire available land area in the arctic. In this case, the evaporation of the oceans becomes less intense as the salinity increases, turning the atmosphere drier (see Figure 3), and consequently allowing less available moisture for the growth of ice glaciers.

For the scenario of $\theta = 0.4$, the time-scale required for sea ice to fill the oceans of the arctic regions suffered a reduction (Figure 4), due to the reduction of ocean area; however, the relative difference between the salinity of 35000 ppm and 260000 ppm remained on the order of 100%. As for the time-scale for ice sheets to cover the land area, there was an increase (Figure 5), not only due to the increase of land area, but also because the less extensive oceans provided less water vapor to the atmosphere through evaporation, causing the average relative humidity to be less than the case of $\theta = 0.7$ (Figure 6). Also, the relative difference of time-scale between 35000 ppm and 260000 ppm remained on the order of 23%. Figures 4, 5 and 6 show the results.
Finally, in the most extreme scenario of $\theta_0=0.1$, the time required to fill the entire arctic land with ice glaciers ranges from 115000 years (salinity=35000 ppm) to 145000 years (salinity=260000 ppm), that is, an increase of around 26% between the maximum and minimum values of salinity. The small ocean area becomes covered in a much shorter time, as expected, but the difference in time-scale between oceans with 35000 ppm and with 260000 ppm remains around 100%. Figures 7 and 8 show the results. This scenario represents a more intense effect of the salinity-evaporation correlation, since the ocean area is smaller and the evaporation rate is proportional to the liquid area exposed to the atmosphere. Consequently, the average atmospheric humidity is lower than the previous cases (Figure 9). The slight increase in the relative difference in land glaciers (26%) compared with the cases of $\theta_0=0.7$ and $\theta_0=0.4$ involves the action of a new phenomenon, which will be explored in the next section.

Although Eq. (6) allows us to work regardless of the total atmospheric pressure at the surface of the planet, since $\phi$ is a dependent variable on the water mass balance, and $p^*$ depends only on the surface temperature, the value of $\alpha$ does depend on the total atmospheric pressure, among other variables not considered in this simulation. For planets with total atmospheric pressure lower than the Earth’s, $\alpha$ will increase, leading to a decrease of the time required to fill the arctic land area with ice sheets ($t_{fill}$).

### 6.2 Shallow oceans

In the simulations performed in the previous subsection, the salinity is also a function of time, since the salt mass in the oceans is constant in the time-scales considered, and the ocean water mass is depleted along time. However, with an average depth of 3.7 km for the previous cases, the volume variation of the temperate and tropical oceans is not enough to increase their salinity to the point of affecting the water vapor saturation pressure, since they are linked with a non-linear relation (see Eq. (7)). Table 7 shows the details of volume and salinity variations for these simulations; it can be seen that ocean volume and salinity both show small variations even for the $\theta_0=0.1$ scenario.
Such periodic changes of orbital elements, in the style of the Croll-Milankovitch cycles of the Earth, are probably common in terrestrial exoplanets, given the available sample of the architecture of multiplanetary systems revealed by exoplanet research. These changes arise essentially from mutual gravitational perturbations between planets, and in exoplanetary systems, if anything, they could well be on average more severe than observed in the Solar System, not only on Earth but also on Mars (Laskar et al. 2004).

\[\begin{array}{cccc}
\theta_0 & 0.1 & 0.2 & 0.3 \\
\text{Fraction of arctic land area} & 0.05 & 0.1 & 0.2 \\
\text{(years)} & 10000 & 50000 & 100000 \\
\end{array}\]

Figure 10 - Evolution of the arctic land area covered by ice glaciers for the cases of \(\theta_0=0.1\) (dashed line), \(\theta_0=0.2\) (dotted line) and \(\theta_0=0.3\) (solid line). Values above 1 mean that the excess ice sheets occupy the temperate region.

These perturbations could well be relevant to climate evolution in low mass exoplanets given the high occurrence of planetary orbits locked in resonances, as well as the higher incidence of sizable eccentricities. Considering the time scales we found in this work, this effect would probably not be very important for Mars, since the periodicity of martian ice-age episodes remains poorly known, controversial, and in the range of 100,000 to 400,000 years (Smith et al. 2016; Schorghofer 2016; Weiss 2019).

The negative feedback effect weakens considerably as the initial ocean coverage (given by \(\theta_0\)) increases. This is explained by the fact that as \(\theta_0\) increases there will be less arctic land area to be filled with ice glaciers; moreover, oceans will have larger water volume, which will feel less the water demand. Therefore, the volume drop in the oceans will be proportionally more modest as compared to that for \(\theta_0=1\). This relation is also translated as a more modest increase in ocean salinities.

Also, for planets with higher values of \(\theta_0\), the steady-state area of land ice sheets becomes more extensive; although our model was not built for land ice sheets penetrating the temperate region, translated by values higher than one for the fraction of arctic land area covered by ice, we deem it robust enough for small extrapolations. Table 9 shows the results for \(\theta_0\) in the range of 0.1 to 0.3, and the same time lapse of 5 x 10^6 years. They indicate that the negative feedback could help delay or stop glaciation for a good range of \(\theta_0\), when acting simultaneously with changes in orbital elements (Croll-Milankovitch-like).

\[\begin{array}{cccc}
\theta_0 & 0.1 & 0.2 & 0.3 \\
\text{Fraction of arctic land area} & 0.05 & 0.1 & 0.2 \\
\text{(years)} & 10000 & 50000 & 100000 \\
\end{array}\]


Table 9 – Final extension of land ice sheets and remaining ocean areas for several values of $\theta_0$ and a time lapse of 500,000 years. Values above 1 for the fraction of arctic land area covered by ice mean that the exice sheets occupy the temperate region.

| $\theta_0$ | % of arctic land area covered by ice | % of planet area covered by ice | % of oceans relative to the original ones | Tropical | Temperate |
|------------|------------------------------------|--------------------------------|-----------------------------------------|----------|----------|
| 0.10       | 0.7                                | 0.0519                         | 0.5                                     | 3.0      | 3.0      |
| 0.15       | 1.0                                | 0.0701                         | 0.5                                     | 3.0      | 3.0      |
| 0.20       | 1.4                                | 0.0924                         | 0.5                                     | 3.0      | 3.0      |
| 0.25       | 1.8                                | 0.1114                         | 0.5                                     | 3.0      | 3.0      |
| 0.30       | 2.25                               | 0.1298                         | 0.5                                     | 3.0      | 3.0      |

6.3 Case study: Mars

The previous simulations were designed for general exoplanets with a range of liquid water coverage in their surface. Moreover, the water coverage was considered the same for each of the three climatic regions. In the case of Mars, we have enough data on the hypothetical ancient ocean that covered the lowlands of the northern hemisphere (Saunders & Schneeberger 1989; Di Achille and Hynek 2010; Parker, Clifford & Parker 2001; Citron, Manga & Hemingway 2019) so that we can afford a customized simulation. Although there is still an ongoing debate as to whether or not the many signs on the surface of the planet are sure imprints of an ocean, or even if the planet ever had enough greenhouse effect to allow liquid water (Ramirez & Craddock 2018), we will assume that it existed in fact. The purpose of the simulation of the putative ocean on Mars is to assess how the salinity-evaporation effect could have delayed the onset of ice ages.

The topography of Mars shows that most of the low terrain is located in the northern hemisphere (with the exception of the Hellas Planitia and Argyre Planitia in the southern hemisphere, whose surface lies well below the topographic datum of Mars); The putative ancient ocean would have covered the entire northern arctic region, and sections of the northern temperate and tropical regions. The onset of an ice age would freeze the waters of the northern arctic and form glaciers in the highlands of the southern arctic regions through the transport of water vapor in the atmosphere. Taking into account that our model does not distinguish the northern from the southern regions (the equinox configuration dictates that both the northern and southern regions receive, on average, the same flux of stellar radiation), the reality of the planet is translated to the model as 50% water coverage for the (entire) arctic region. Following the same reasoning for the temperate and tropical regions, and considering the estimated water extent of the ancient northern ocean and of water bodies in Hellas and Argyre basins, we estimated a water extent of 35% and 30% for the temperate and tropical regions respectively. This is equivalent to an overall surface water fraction of 34.25%, very similar to the estimate of 35.7% made by Di Achille and Hynek (2010), and a more conservative one.

As for the total water volume, Di Achille and Hynek (2010) have estimated a value of $1.24 \times 10^8$ km$^3$, taking into account the northern ocean and the areas of Hellas and Argyre basins; however, they admit the possibility of overestimation due to the addition of ancient paleolakes in craters which might have been formed later. The volume of the Arabia ocean estimated by Citron, Manga & Hemingway (2019) is $5.5 \times 10^7$ km$^3$. So, we chose the intermediate value of $9 \times 10^7$ km$^3$ for our base case, which translates to an average water depth of 1.82 km, which is higher than our previous definition of shallow ocean, but lower than the average depth of 3.7 km used in the initial simulations.

The results of the simulations are shown in Figures 11 to 13 and in Table 10. As expected, the results for the arctic ocean are similar to the case of $\theta_0 = 0.7$ for general planets with deep oceans, since for the martian arctic region $\theta_0 = 0.5$. As for the evolution of the land glaciers, the result is intermediate between the results for $\theta_0 = 0.7$ and $\theta_0 = 0.4$ for deep oceans, indicating an increase of ~23% in the time required to fill the arctic region, and considering the usual salinity range of 35000 to 260000. Table 10 shows that although the initial average depth of 1.82 km implies a stronger negative feedback due to increased salinity and decreased ocean area, when compared with the average depth of 3.7 km, it is still not strong enough to decrease the growth rate of land ice sheets considerably over time. Considering the much longer periodicity of martian ice ages (~10$^5$ to a few times 10$^6$ years) as compared to that of terrestrial ice ages (about 40,000 years for the obliquity cycle), it is unlikely that the negative feedback due to salinity in putative martian oceans could have made a difference, as compared to much longer, and presumably stronger, climatological cycles on Mars, unless the average depth of such putative oceans was much less than the estimates used for the model. This possibility will be explored in future work.

Figure 11 - Evolution of the fraction of martian arctic ocean area covered by ice, for 3 different ocean salinities.
is equivalence equation, only an atmosphere rich in SO₄;







\[ (iM)_1 = (iM)_2 \]  \hspace{1cm} (15)

For the units used in this work (ppm of solute), this equivalence can be written as

\[ \frac{i_1}{MW_1} \left[ \frac{ppm_i}{1000} \right] = \frac{i_2}{MW_2} \left[ \frac{ppm_i}{1000} \right] \]  \hspace{1cm} (16)

where the subscript 1 refers to NaCl and the subscript 2 refers to MgSO₄. Table 11 shows the equivalent concentration of MgSO₄ for the initial concentrations of NaCl used in the simulations.

| Initial Conc. of NaCl (ppm) | Equivalent Conc. of MgSO₄ (ppm) | Observation |
|-------------------------------|---------------------------------|-------------|
| 35000                         | 80661                           | Beyond MgSO₄ solubility |
| 150000                        |                                 | Beyond MgSO₄ solubility |
| 260000                        |                                 |                 |

Table 11 – Conversion of concentrations with the same colligative effect

In fact, the reader who might be interested in substituting the NaCl in the previous simulations with other salts may use this equivalence equation, only taking care to check if the resulting concentration is higher than de value of the solubility in water.

6.4 Astrobiological implications

Planets smaller than the Earth are expected to have high salt contents in their oceans, due to the faster planetary cooling process, which would have slowed down or even stopped plate tectonics and its

The conjecture that its ancient ocean and lakes might have been rich in sulfate salts, due to the interaction between the water and a sulfur-rich atmosphere (SO₂, SO₃). However, the average solubility of sodium and calcium sulfate in water at the temperatures of interest (0 to -30 C) is well below the one of sodium chloride (Perry 1997 b). Sulfide and sulfite salts have even lower solubilities. For calcium sulfate, the solubility ranges from 0.21 to 0.24 g per 100 ml of water, depending on the hydration degree, and for sodium sulfate the value ranges from 19 to 44 g per 100 ml of water at 20 C.

Further, since the molecular weight of sulfite and sulfate salts are much higher than that of NaCl, their molality, which drives the colligative effects under study in this work, would be very small. As for magnesium sulfate, which might have been a salt dissolved in Mars’ waters, the solubility even surpasses the one for NaCl (Perry 1997 b), leaving only the molecular weight as a hindrance. The hypothesis of sulfate salt dissolved in water on Mars presents other complications. For example, if its origin was an atmosphere rich in SO₂ and SO₃, its massive dissolution in water would have altered its pH, altering as a consequence the solubility of salts.

It is not our intention to exhaust the ongoing debate about the complex sulfate problem on Mars (King, Lescinsky & Nesbitt 2004; Chow & Seal II 2007; Marion et al. 2013; Kaplan et al. 2016), or even suggest a new hypothesis. We have performed the simulations with NaCl as before, based on the indication that the mantle of Mars is rich in Na and Cl, and poor in S (Yoshizaki & McDonough 2020), but this time we have also provided the reader with a colligative effect equivalent, that is, for the initial value of NaCl concentration in ppm, we provided the equivalent value in ppm of MgSO₄, using the colligative equivalence equation

\[ (iM)_1 = (iM)_2 \]  \hspace{1cm} (15)

We maintained our assumption of an ocean with mainly NaCl salt with three different initial concentrations. Mars apparently did not have a plate tectonics process like the one which exists in our planet (Nicola & Padovan 2019) and probably no hydrothermal systems as well, so its putative ancient ocean, only accumulated salt due to weathering of dry land; therefore, the salt content of the ocean was probably high during most of its existence. The discovery of sulfate deposits on Mars (Klingelhöfer et al. 2004, Rieder et al. 2004) leads to the conjecture that its ancient ocean and lakes might have been rich in sulfate salts, perhaps due to the interaction between the water and a sulfur-rich atmosphere (SO₂, SO₃). However, the average solubility of sodium and calcium sulfate in water at

![Figure 12 - Evolution of the fraction of martian arctic land area covered by ice, for 3 different ocean salinities](image1.png)

![Figure 13 - Evolution of martian global relative humidity, for 3 different ocean salinities](image2.png)

| Table 10 – Evolution of salinities and ocean volumes (10⁶ km²) for the three different initial salinities for the Mars simulations |
|--------------------------------------------------|--------------------------------------------------|--------------------------------------------------|
| Initial salinity | Initial salinity | Initial salinity |
| 35000 ppm | 15000 ppm | 260000 ppm |
| Final Vol. (temp.) | 4.039 | 4.054 | 4.072 |
| Final Vol. (trop.) | 2.459 | 2.485 | 2.515 |
| Final sal (temp.) | 36518 | 155210 | 266970 |
| Final sal (trop.) | 39228 | 164280 | 27890 |
| Initial Vol. (temp.) | 4.221 | 4.221 | 4.221 |
| Initial Vol. (trop.) | 2.769 | 2.769 | 2.769 |
associated hydrothermal system. These intrinsically more saline (deep) oceans will resist the glaciation process by slowing down the growth rate of land ice sheets and glaciers. The time extension could be up to around 23%, considering the maximum salinity range studied, between Earth’s and the near saturation salinity. This effect could possibly counteract Croll-Milankovitch-like orbital forcings that push the planetary climate system to ice ages. From the Gaian point of view (Lenton 1998; Nicholson et al. 2018), this stabilizing phenomenon can be viewed as an abiotic mechanism to avoid, or at least mitigate, ice ages potentially disastrous to life, since smaller planets would have less massive oceans, with less heat capacitance, and therefore more susceptible to total freezing (iceball events) than the oceans of larger planets.

The appearance of life depends upon many physical factors (Lingam & Loeb 2019), not to mention biochemical factors, but it is thought that liquid water is a necessary requirement, as the definition of the Habitable Zone attests. Therefore, if we assume that life in other planets is born preferentially in oceans and spends most of its existence there, before migrating to dry land, as has happened on Earth, the salinity-evaporation effect could be viewed as an extra protection mechanism.

For planets with shallow oceans, that is, the ones with the highest susceptibility to disastrous ice ages, the intrinsic salinity-evaporation effect is amplified with time, due to the process of glaciation, acting simultaneously with the reduction of ocean surface area, in a dynamic negative feedback process that reinforces the opposition to other forcings, mitigating or even effectively stopping the glaciation process. Besides the danger to life mentioned above, intense ice ages can also have other deleterious consequences to life productivity, like an excessive reduction of ocean coverage fraction (Lingam & Loeb 2019 b).

The marine life of such oceans would have to develop adaptation strategies in order to withstand high salinity variations on time scales of $10^5–10^7$ years. This is, in principle, a viable evolutionary scenario, since on Earth there are extremophiles which thrive in brines, beyond life in the oceans.

Once the glaciation process is reverted, due to a number of possible factors, the return of land ice to the oceans as liquid water will restore a lower level of salinity, and a more extensive ocean area, which is able to evaporate more easily, increasing the greenhouse effect that had already been increased in order to end the ice age. That is, the salinity-evaporation effect, in shallow oceans, is also a destabilizing (positive feedback) agent regarding planet warming. But, considering also that smaller planets are more susceptible to cooling, due to a probably thinner atmosphere, this destabilizing factor may also turn out to be key to life survival, for the likelier planetary habitability bottleneck would tend to be excessively cold climate, and not excessively hot ones. If so, the salinity-evaporation effect could be viewed as an abiotic climatic self-regulation, always aimed at keeping the planet away from excessive glaciation, although the name “cycle” would possibly not be a proper one, as contrasted to the carbonate-silicate cycle (Kasting 2010; Rushby 2018), which acts as a negative feedback for both high and low temperatures.

It is interesting to note that, as long as greenhouse effect goes, the salinity-evaporation effect also acts as a destabilizing (positive feedback) agent on glaciation, for the lower drier atmosphere, caused by the depleted evaporation, helps the decrease of surface temperatures. Although our model is not rigorous with respect to the greenhouse effect caused by water vapor, we did take it into account, and we think that the salinity-evaporation effect (negative feedback) on the growth rate of ice glaciers outweighs the humidity factor, either by intrinsically high salinity or by the coupling of high salinity with the feedback of shallow oceans. However, more complete simulations are needed to assess more precisely the relative forces of these agents, which we intend to pursue in a forthcoming paper.

The increased retention of water in the liquid state provided by the salinity-evaporation relationship could also delay the volatile loss to space in planets subject to high EUV/X-rays flux from the parent star. This phenomenon has not been modeled in the present work, which moreover deals with time-scales of $10^5–10^7$ years, while volatile loss is a process with time-scales of $10^9$ years. Yet the influence of salinity on volatile loss would be a logical consequence, since a drier atmosphere would convey a smaller amount of volatiles to its upper region, where the molecules are split by radiation and particle fluxes, and lost to space. Quantitative studies should be performed in order to assess this possibility, and it is our intention to do so in the near future.

7 CONCLUSIONS

Our climate toy model, which captures the essentials of ocean evaporation and ice formation dynamics, coupled with radiative equilibrium temperatures, indicates clearly that the salinity-evaporation and salinity-freeze point correlations, in small rocky planets with intrinsically high salinities and a low volatile content, act as a climatic buffer in planets entering ice ages, due to Croll-Milankovitch-like orbital evolution, through a considerable decrease in land ice sheet growth rate, up to ~ 23% considering the maximum salinity range.

Moreover, for planets with shallow oceans (with average depth in the order of hundreds of meters), the self-reinforcement of the salinity effect, simultaneously with the reduction of ocean area, act as a negative feedback on the glaciation process, to the point of decreasing and even effectively stopping the land ice sheet growth rate during the time-scale of the simulations.

These two phenomena (intrinsically higher salinity, and the negative feedback in the case of shallow oceans) would be considered, under the point of view of astrobiology, as protection mechanisms against glaciations in small, volatile-poor rocky planets: such glaciations would be more damaging to life than in the case of Earth-like planets, with vast and deep oceans.

The enhanced time scale of ice formation in planets with shallow oceans is comparable to the time-scale of Croll-Milankovitch-like changes of orbital elements, which leads the Earth system to ice
ages. Therefore, it is possible that the salinity-evaporation feedback effect on small planets with shallow oceans (or Earth sized planets deficient in water) may delay ice ages or even prevent them from developing altogether.

The assumed high salinities employed are expected for habitable planets considered in this study, that is, planets with sizes comparable to that of the Earth and smaller, where the plate tectonics process is probably less intense, and, consequently, where hydrothermal systems operate less effectively in removing salt from oceans.

The results are a function of, mainly, the salinity of the oceans and the fraction of the planet area covered by oceans. In the case of shallow oceans, the negative feedback effect increases the inertia of the planetary system, as the glaciation process evolves.

The salinity-evaporation correlation, reinforced with the dynamic negative feedback in the case of shallow oceans, can be understood as an abiotic climatic self-regulation cycle, that keeps small planets, with possess weak or no plate tectonics and small amounts of volatiles, away from extremes of glaciation, which could otherwise eventually freeze all oceans to the bottom and kill all marine life, since the heat capacitance of such oceans would not match those of Earth-like planets, with vast and deep oceans. Furthermore, these phenomena may help delay or even keep in check the loss of volatiles to space for small habitable planets.

The salinity-evaporation effect, along with its reinforcement as a negative feedback in planets with shallow oceans, given the likely prevalence of small, volatile-poor bodies in the rocky planet population, are probably a common phenomenon in rocky exoplanets in the Habitable Zone, and explored for the first time in the present paper, and we suggest that both effects should thereof be incorporated in future climatic models, and in astrobiological considerations regarding physical conditions that could enhance the resilience of life in the Universe.

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REFERENCES

Adams E. R., Seager S., Elkins-Tanton L., 2008, ApJ, 673, 1160
Airapetian V. S., Glocer A., Khazanov G. V., Loyd R. O. P., France K., Sojka J., Danchi W. C., Liemohn M. W., 2017, ApJL, 836, L3
Anglada-Escudé G. et al., 2016, Nature, 536, 437
Babkin V. I., 2019, Evaporation from the surface of the Globe; in Shiklomanov I. A., ed, Hydrological Cycle, UNESCO-EOLSS
Barrie W. J., 2008, Int. J. Astrobiology, 7, 143
Bolmont E., Selsis F., Owen J. E., Ribas I., Raymond S. N., Leconte J., Gillon M., 2017, MNRAS, 464, 3728
Buck A. L., 1981, J. Appl. Meteor., 20, 1527
Chow J-Ming, Seal II R. R., 2007, Journal of Geophysical Research, 112, E11004
Citron R. I., Manga M., Hemingway D. J., 2018, Nature, 555, 643
Clifford S. M., Parker T. J., 2001, Icarus, 154, 40
D’Angelo M., Casaux S., Kamp I., Thi W. –F., Woitke P., 2019, A&A, 622, A208
Di Achille G., Hynek B. M., 2010, Nat. Geosci., 3, 459
do Nascimento Jr. J. D. et al., 2016, ApJL, 820, L15
Dong C., Huang Z., Lingam M., Toth G., Gombosi T., Bhattacharjee A., 2017, ApJ Letters, 847, L4
Elkins-Tanton, L.T., 2012, Ann. Rev. Earth Planet. Sci. 40, 113
Franck S., Block A., von Bloh W., Bounama C., Schellnhuber H. –J., Svirezhev Y., 2000, PSS, 48, 1099
Genda H., Ikoma M. 2008, Icarus, 194, 42
Harberge Jr. G. E., 1955, The effect of salinity on evaporation, in Studies of Evaporation, Geological Survey Professional Paper 272
Henshaw P. C., Charlson R. J., Burges S. J., 2000, in Jacobson M. C., Charlson R. J., Rodhe R., Orians G. H., eds, Earth System Science, Academic Press, International Geophysics Series, v. 72, San Diego, p. 112
Hovland M., Rueslåttén H., Johnsen H. K., 2018, Marine and Petroleum Geology, 92, 128, 2018
Ida S., Yamamura T., Okuzumi S., 2019, A&A, 624, A28
Kaltenegger L., Sasselov D., 2013, EGU Gen. Assembly, 14141
Kaplan H.H., Milíken R. E., Fernández-Remolar D., Amils R., Robertson K., Knoll A. H., 2016, Icarus, 275, 45
Kasting J., 2010, How to find a Habitable Planet, Princeton University Press, Princeton New Jersey, p. 49
Kasting J. F., Whitmire D. P., Reynolds R. T., 1993, Icarus, 101, 108
Kiefer H. H., Jakosky B. M., Snyder C. W., 1992, The Planet Mars: From antiquity to the present. In Kiefer H. H., Jakosky B. M., Snyder C. W., Matthews M. S., eds, Mars, Univ. of Arizona Press, Tucson, p. 1-33
King P. L., Lescinsky D. T., Nesbitt H. W., 2004, Geochimica et Cosmochimica Acta, 68, 4993
Kitzmann D. et al. 2015, MNRAS, 452, 3752
Klingelhöfer G. et al., 2004, Science, 306, 1740
Kodama T., Genda H., O’ishi R., Abe-Ouchi A., Abe Y., 2019, J. Geophy. Res.: Planets, 124, 2306
Kopparapu R. K. et al., 2013, ApJ, 765, 131
Lammer H. et al., 2009, A&A Rev., 17, 181
Landais F., Schmidt F., Lovejoy S., 2019, MNRAS, 484, 787
Langmuir C. W., 1955, PSS, 48, 1099
Langmuir C. W., Broecker W., 2012, How to build a habitable planet. Princeton Univ. Press, Princeton, NJ
Laskar J., Correia A. C. M., Gastineau F., Joutel F., Levrard B., Robutel P., 2004, Icarus, 170, 343
Launius R. D., 2012, Life, 2, 255
Lebrun, T., Massol, H., Chassefière, E., Davaille A., Marcq E., Sarda P., Leblanc F., Brandeis G., 2013, J. Geophys. Res.: Planets, 118, 1155
Mayor M., Queloz D., 1995, Nat 378, 355
Lenton T. M., 1998, Nature, 394, 439
Lingam M., Loeb A., 2019, Rev. Mod. Phys., 91, 021002
Lingam M., Loeb A., 2019b, AJ, 157, 25
Lissauer J. J., 2007, ApJ, 660, 149
Marion G. M., Kargel J. S., Crowley J. K., Catling D. C., 2013, Icarus, 225, 342
Molaverdikhani K., Helling C., Lew B. W. P., MacDonald R. J., Samra D., Iro N., Woitke P., Parmentier V., 2020, eprint arXiv:2001.03668
Morbidelli A., Lunine J. I., O’Brien D. P., Raymond S. N., Morishima R., Walsh K. J., 2014, in Beuther H., Klessen R. S., C. P. Dullemond, eds, Protostars and Planets VI, Univ. Arizona Press, Tucson, p.835
Wallack N. L. et al., 2019, AJ, 158, 217
Weiss D. K., 2019, Earth and Planetary Science Letters, 528, 115845
West A. A., Hawley S. L., Bochanski J. J., Covey K. R., Reid N., Dhillon S., Hilton E. J., Masuda M., 2008, AJ, 135, 785
Wood B. E., Müller H. –R., Zank G. P., Linsky J. L., 2002, ApJ, 574, 412
Wood B. E., Müller H. –R., Zank G. P., Linsky J. L., Refield S., 2005, ApJ, 628, L143
Yoshizaki T., McDonough W. F., 2020, Geochimica et Cosmochimica Acta, 273, 137
Zuluaga J. I., Bustamante S., Cuartas P. A., Hoyos J. H., 2013, ApJ, 770, 23

van Dishoeck E. F., Bergin E. A., Lis D. C., 2014, in Beuther H., Klessen R. S., C. P. Dullemond, eds, Protostars and Planets VI, Univ. Arizona Press, Tucson, p.835
Wallack N. L. et al., 2019, AJ, 158, 217
Weiss D. K., 2019, Earth and Planetary Science Letters, 528, 115845
West A. A., Hawley S. L., Bochanski J. J., Covey K. R., Reid N., Dhillon S., Hilton E. J., Masuda M., 2008, AJ, 135, 785
Wood B. E., Müller H. –R., Zank G. P., Linsky J. L., 2002, ApJ, 574, 412
Wood B. E., Müller H. –R., Zank G. P., Linsky J. L., Refield S., 2005, ApJ, 628, L143
Yoshizaki T., McDonough W. F., 2020, Geochimica et Cosmochimica Acta, 273, 137
Zuluaga J. I., Bustamante S., Cuartas P. A., Hoyos J. H., 2013, ApJ, 770, 23