Oxygen False Positives on Habitable Zone Planets Around Sun-Like Stars

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Abstract  Oxygen is a promising exoplanet biosignature due to the evolutionary advantage conferred by harnessing starlight for photosynthesis, and the apparent low likelihood of maintaining oxygen-rich atmospheres without life. Hypothetical scenarios have been proposed for non-biological oxygen accumulation on planets around late M-dwarfs, where the extended pre-main sequence may favor abiotic O2 accumulation. In contrast, abiotic oxygen accumulation on planets around F, G, and K-type stars is seemingly less likely, provided they possess substantial non-condensable gas inventories. The comparative robustness of oxygen biosignatures around larger stars has motivated plans for next-generation telescopes capable of oxygen detection on planets around sun-like stars. However, the general tendency of terrestrial planets to develop oxygen-rich atmospheres across a broad range of initial conditions and evolutionary scenarios has not been explored. Here, we use a coupled thermal-geochemical-climate model of terrestrial planet evolution to illustrate three scenarios whereby significant abiotic oxygen can accumulate around sun-like stars, even when significant noncondensable gas inventories are present. For Earth-mass planets, we find abiotic oxygen can accumulate to modern levels if (1) the CO2:H2O ratio of the initial volatile inventory is high, (2) the initial water inventory exceeds ~50 Earth oceans, or (3) the initial water inventory is very low (<0.3 Earth oceans). Fortunately, these three abiotic oxygen scenarios could be distinguished from biological oxygen with observations of other atmospheric constituents or characterizing the planetary surface. This highlights the need for broadly capable next-generation telescopes that are equipped to constrain surface water inventories via time-resolved photometry and search for temporal biosignatures or disequilibrium combination biosignatures to assess whether oxygen is biogenic.

Plain Language Summary  Next-generation telescopes will search for life on exoplanets by looking for the spectral signatures of biogenic gases. Oxygen has been considered a reliable biosignature gas, especially for planets around sun-like stars where non-biological, photochemical production is unlikely. This motivates plans for future telescopes specifically designed for oxygen detection. Here, we develop a coupled model of the atmosphere-interior evolution of terrestrial planets to show that lifeless planets in the habitable zone could develop oxygen-rich atmospheres relatively easily. These false positives for biological oxygen could be distinguished from inhabited planets using other contextual clues, but their existence implies next-generation telescopes need to be capable of characterizing planetary environments and searching for multiple lines of evidence for life, not merely oxygen.

1. Introduction

The search for life beyond Earth is a key motivator for exoplanet astronomy. Among the various life detection approaches that have been proposed, atmospheric oxygen is arguably the most promising biosignature. This is because any organism that adapts to exploit free energy from starlight will have a competitive advantage over organisms that are limited by geochemical sources of free energy. The earliest incontrovertible evidence for life on Earth around 3.5 Ga coincides with fossilized photosynthetic stromatolites (Buick, 2008). While it is unknown whether the evolution of oxygenic photosynthesis is contingent—the metabolism emerged only once in Earth's history (Mulkidjianian et al., 2006)—oxygenic photosynthesis confers a unique evolutionary advantage over other forms of photosynthesis since the required substrates, carbon dioxide...
(CO₂) and water (H₂O) are likely ubiquitous on habitable worlds. Moreover, the accumulation of biogenic oxygen in planetary atmospheres, as has occurred on Earth (Holland, 2006; Lyons et al., 2014), is readily detectable over interstellar distances thanks to absorption features in the visible, near IR, and (for ozone) thermal infrared (reviewed by Meadows et al., 2018). Indeed, three of the four proposed mission concepts currently under consideration by the 2020 Astrophysics Decadal Survey are specifically designed to be capable of life detection via oxygen or ozone (Fischer et al., 2019; Gaudí et al., 2020; Meixner et al., 2019).

Ground-based Extremely Large Telescopes may also be capable of oxygen detection via high resolution spectroscopy (Leung et al., 2020; Rodler & López-Morales, 2014; Snellen et al., 2013). The James Webb Space Telescope (JWST) could conceivably detect oxygen/ozone biosignatures for nearby transiting planets, although this would probably require a prohibitively large number of transits for known targets (Fauchez et al., 2020; Krissansen-Totton, Garland, et al., 2018; Lustig-Yaeger et al., 2019; Wunderlich et al., 2019).

Finally, oxygen is often considered to be a good biosignature because—apart from the exceptions discussed below—it is seemingly difficult for habitable zone terrestrial planets to maintain oxygen-rich atmospheres without life (reviewed by Meadows et al., 2018).

The terrestrial planets that will be accessible to spectroscopic characterization with JWST and ground-based ELTs will orbit M-dwarfs due to the favorable signal-to-noise of M-dwarf transits. This is unfortunate from the standpoint of recognizing oxygen biosignatures because several features of M-dwarfs make them susceptible to non-biological oxygen accumulation. In particular, the extended pre-main sequence of late M-dwarfs could yield habitable zone terrestrial planets with hundreds or thousands of bar O₂ from XUV-driven hydrogen loss (Luger & Barnes, 2015). At least some of this oxygen will likely dissolve in a surface magma ocean and be sequestered in the mantle, but retaining oxygen-rich atmospheres is still possible, especially for highly irradiated terrestrial planets (Barth et al., 2020; Schaefer et al., 2016; Wordsworth et al., 2018). The early development of abiotic, oxygen-rich atmospheres could prevent the subsequent emergence of life by precluding prebiotic chemistry (Wordsworth et al., 2018). It has also been argued that the spectral energy distributions of M-dwarfs are favorable for the photochemical accumulation of O₂ and O₃ (Domagal-Goldman et al., 2014; Gao et al., 2015; Harman et al., 2015; Tian et al., 2014). While recent assessments of the role of lightning in such photochemical models (Harman et al., 2018) and improved near-UV water cross sections (Ranjan et al., 2020) may preclude some of these scenarios, photochemical runaways yielding O₂-CO rich atmospheres remain a strong possibility for late M dwarfs (Hu et al., 2020).

More fundamentally, however, whether M-dwarf terrestrial planets can sustain habitable surface conditions, or indeed any kind of substantial atmosphere, for billions of years given the harsh stellar radiation environment is unknown (Airapetian et al., 2017; Dong et al., 2018; Shields et al., 2016; Zahnle & Catling, 2017). Premature claims of biogenic oxygen on M-dwarf planets will rightfully be met with skepticism.

These challenges to M-dwarf habitability and potential ambiguities with oxygen biosignatures motivates the need to characterize habitable zone planets around F, G, and K-type stars (Arney et al., 2019; National Academies of Sciences & Medicine, 2018). The relatively short pre-main sequence of larger stars mean that the accumulation of non-biological oxygen is less likely. Indeed, the most plausible mechanism for non-biological O₂ accumulation on habitable zone planets around sun-like stars is due to small atmospheric inventories of non-condensable species like N₂; this weakens the water cold trap and may produce high stratospheric water abundances and therefore elevated H escape rates (Kleinböhl et al., 2018; Wordsworth & Pierrehumbert, 2014). Whether terrestrial planets are likely to form with low nitrogen atmospheric inventories is unknown, but in principle, it might be possible to rule out low non-condensable scenarios by looking for the spectral signatures of abundant nitrogen (Schwieterman et al., 2015) or via inferring background gas abundances from retrieved surface pressure (Feng et al., 2018).

The broader question as to whether habitable zone terrestrial planet evolution tends toward anoxic atmospheres, or whether non-biological oxygen production via water photodissociation and hydrogen escape can overwhelm geochemical sinks, has not been explored. The robustness of oxygen biosignatures rests on the assumption that for temperate planets with effective cold traps, small abiotic oxygen source fluxes from H escape will be overwhelmed by geological sinks. To test this assumption, it is necessary to model the redox evolution of terrestrial planets from formation onwards. This is because planetary redox evolution depends on both the initial state of the atmosphere and mantle after the magma ocean has solidified, and on the subsequent internal evolution and atmospheric state. Interior evolution dictates crustal production rates...
and outgassing fluxes, which determine the efficiency of geologic sinks of oxygen. Moreover, self-consistent modeling of surface climate and weathering processes is necessary because these modulate surface volatile inventories and oxygen production via atmospheric H escape.

Here, we explore the tendency of terrestrial planets to produce abiotic, oxygen-rich atmospheres with a fully coupled model of planetary redox, climate, and thermal evolution. Given the vagaries of planetary accretion, migration, and core formation, it is likely that habitable zone terrestrial planets exist with a broad range of initial volatile inventories (Raymond et al., 2004, 2013; Righter, 2015). Indeed, both planet formation models (Raymond et al., 2004) and exoplanet demographics suggest that water-rich terrestrial planets are abundant (Zeng et al., 2019), and several candidate waterworlds are amenable to JWST characterization (Bennike et al., 2019; Grimm et al., 2018). In contrast, transit and thermal phase curve observations show that the terrestrial exoplanet LHS 3844b is either a bare rock or possesses a thin, high mean-molecular weight atmosphere susceptible to stellar wind erosion (Diamond-Lowe et al., 2020; Kreidberg et al., 2019), and modeling implies it must have formed with less volatiles than the Earth (Kane et al., 2020). Relative abundances of volatiles may also be variable; C/H mass ratios in carbonaceous chondrites vary from ~1.6 to 6.4 (Nittler et al., 2004). We thus apply our model to explore the possibility of non-biological oxygen accumulation for a range of initial volatile inventories. Crucially, we avoid varying background N abundances because our focus is on investigating oxygen false positives on planets with comparatively high background non-condensables inventories, that is, sufficient to maintain a cold trap under temperate conditions.

2. Materials and Methods

Our model is summarized schematically in Figure 1. A complete description is provided in the supporting information and the Python code is open source (https://doi.org/10.5281/zenodo.4539040). Here, we summarize the salient features of the model necessary for interpreting key findings. There are no biological
sources and sinks of volatiles because we are testing the capacity of lifeless planets to accumulate oxygen. Unless stated otherwise, all calculations assume an Earth-mass planet at 1 AU.

Planetary evolution is divided into an initial magma ocean phase, and a subsequent solid-mantle phase, as shown in Figure 1, although a planet may transition between magma ocean and solid mantle multiple times. The model is initialized with a fully molten mantle and some endowment of volatiles, radiogenic inventory, and an initial mantle oxygen fugacity (i.e., after core formation, Figure 1, left). Magma ocean solidification follows previous models such as Lebrun et al. (2013) and Schaefer et al. (2016): the magma ocean freezes from the core, upwards, as governed by the following equation:

\[
V_{\text{mantle}} \rho_m Q_{\text{radioactive}} = 4 \pi q_m r_p^2 + Q_{\text{core}} + 4 \pi \rho_m H_{\text{fusion}} r_s \frac{dr_s}{dt} = V_{\text{mantle}} \rho_m r_p \frac{dT_p}{dt}
\]

Here, \( V_{\text{mantle}} \) is the volume of the molten mantle, \( \rho_m \) is the average density of the mantle, \( Q_{\text{radioactive}} \) is radiogenic heat production per unit mass, \( r_p \) is planetary radius, \( H_{\text{fusion}} \) is the latent heat of fusion of silicates, \( r_s \) is the solidification radius, \( c_p \) is the specific heat of silicates, \( Q_{\text{core}} \) is the heatflow from the metallic core, and \( T_p \) is mantle potential temperature. The heatflow from the interior, \( q_m \), is calculated using 1-D convective parameterization, with temperature-dependent magma ocean viscosity, \( \nu(T_p) \), (see Figure S3 and supporting information Section A.3):

\[
q_m = C_{qm} \left( \frac{T_p - T_{\text{surf}}}{\nu(T_p)} \right)^{1/3}
\]

Here, \( T_{\text{surf}} \) is mean surface temperature, and \( C_{qm} \) is a constant that depends on thermal conductivity, thermal diffusivity, critical Rayleigh number, gravity, and thermal expansivity. Supporting information Section A.1 describes this convection parameterization in more detail. Equation 1 continues to govern the thermal evolution of the mantle after the magma ocean has solidified with \( dr_s/dt = 0.0 \).

During magma ocean solidification H, C, and O are partitioned between dissolved melt phases, crystalline phases, and the atmosphere by assuming chemical equilibrium (supporting information Section A.7). The rate at which the mantle freezes is controlled by outgoing longwave radiative (OLR), which is balanced by heat from interior and absorbed shortwave radiation (ASR) from the host star at every timestep (see climate model description below):

\[
q_m + \text{ASR} = \text{OLR}
\]

During the magma ocean phase, planetary oxidation may occur from the loss of hydrogen to space (less oxygen drag):

\[
\text{H}_2\text{O} \rightarrow \text{H}_2(\text{space}) + 0.5\text{O}_2
\]
OLR in the presence of condensable water vapor, a dry adiabat to moist adiabat to isothermal atmospheric structure is assumed (Kasting, 1988). To calculate ASR across a wide range of temperatures, we adapted the albedo parameterization described in Pluriel et al. (2019). Refer to supporting information Section A.5 for full details of radiative transfer calculations along with example outputs.

When heat from accretion and short-lived radiogenics is sufficiently dissipated—the timescale for which is controlled by insolation and greenhouse warming from outgassed volatiles—a planet’s surface temperature may drop below the solidus and the magma ocean phase is over. At this point, the model transitions to solid-state mantle convection and temperate geochemical cycling (Figure 1, right). The redox budget during solid-state evolution is modeled as follows: the only source of oxygen is still atmospheric escape using the same parameterization as described above. However, there are now numerous crustal sinks for oxygen including (i) subaerial and submarine outgassing of reduced species (e.g., H, CO, and CH₄), (ii) water-rock serpentinizing reactions that generate H₂ (the “wet crustal” sink), and (iii) direct oxidation of surface crust by atmospheric oxygen (the “dry crustal” sink):

\[
\begin{align*}
&\text{(i), } \text{H}_2 + 0.5\text{O}_2 \rightarrow \text{H}_2\text{O} \\
&\text{(ii), } \text{CO} + 0.5\text{O}_2 \rightarrow \text{CO}_2 \\
&\text{(iii), } 2\text{FeO} + 0.5\text{O}_2 \rightarrow \text{H}_2 + 2\text{FeO}_{1.5}
\end{align*}
\]

The sizes of these three oxygen sinks are self-consistently calculated from the planetary interior evolution and mantle volatile content. Outgassing fluxes are calculated using the melt-gas equilibrium outgassing model of Wogan et al. (2020); outgassing fluxes depend on mantle oxygen fugacity, degassing overburden pressure, the volatile content of the mantle, specifically H₂O and CO₂ content, and the rate at which melt (new crust) is produced (see supporting information Section A.10). The possible influence of pressure overburden on redox evolution has been discussed previously (Wordsworth et al., 2018), and is quantifiable within our outgassing model framework. Dry and wet crustal sinks for oxygen similarly depend on crustal production rates, and are described in full in supporting information Sections A.12 and A.13. Crustal production is calculated from interior heatflow, which is modulated by temperature-dependent mantle viscosity and radiogenic heat production (supporting information Section A.10). We assume plate tectonics when calculating crustal production rates to maximize crustal sinks of oxygen.

Weathering processes (Krissansen-Totton, Arney, et al., 2018) and the deep hydrological cycle (Schaefer & Sasselov, 2015) are explicitly modeled because climate and surface volatile inventories control crustal oxygen sinks and atmospheric escape processes. Our model incorporates a rudimentary carbon cycle, which is a simplified version of that described in Krissansen-Totton, Arney, et al. (2018). Carbon is added to the atmosphere via magmatic outgassing (described above), and returned via continental weathering and seafloor weathering, whose relative contributions depend on climate and the total surface water inventory. Our carbon cycle parameterization is described in supporting information Section A.11 and the deep hydrological cycle in supporting information Section A.12. Self-consistent climate modeling enables an assessment of whether abiotic oxygen can coexist with a habitable surface climate, which would be especially problematic for unambiguous biosignature gas interpretations.

The time evolution of volatile reservoirs during both the magma ocean phase and the solid state evolution is governed by the following equations (see supporting information Section A.6 for details):

\[
\begin{align*}
&\frac{dM_{\text{solid}}}{dt} = 4\pi\rho_m k_{\Lambda_i} f_{\Lambda_i} r_s^2 \frac{dr_s}{dt} + F_{\text{ingas}} - F_{\text{outgas}} \\
&\frac{dM_{\text{fluid}}}{dt} = -4\pi\rho_m k_{\Lambda_i} f_{\Lambda_i} r_s^2 \frac{dr_s}{dt} + F_{\text{ingas}} + F_{\text{outgas}} - \text{Esc}_{\Lambda_i}
\end{align*}
\]

Here, \(\Lambda_i\) represents a generic volatile species (e.g., H₂O, CO₂, free O). The first term on the right hand side represents the transfer of volatiles from the fluid phases (magma ocean + atmosphere) to the solid mantle as the magma ocean solidifies: \(k_{\Lambda_i}\) is the melt-solid partition coefficient for species \(\Lambda_i\), and \(f_{\Lambda_i}\) represents...
the mass fraction of the volatile species in the magma. The remaining fluxes are subaerial plus submarine outgassing from the mantle to the atmosphere, $F_{\text{outgas}}$, ingassing from the atmosphere to the mantle (e.g., crustal oxidation or hydration), $F_{\text{ingas}}$, and escape to space $E_{\text{Esc}}$. Because we are assuming a plate tectonics regime, the model does not separately track volatile reservoirs in the crust and mantle. Instead, we assume carbon, oxygen, and water added to crust is immediately subducted into the mantle; a single, well-mixed “interior” reservoir is used to represent storage of volatiles in solid silicates. The extremely broad range of crustal hydration and crustal oxidation efficiency factors sampled (see below) can accommodate differing subduction and arc volcanism efficiencies.

Note that the model only tracks C, H, and O-bearing species, as well as Fe$^{2+}$/Fe$^{3+}$ speciation in the interior. Nitrogen fluxes are not modeled, and we instead assumed a 1 bar N$_2$ background partial pressure in all model runs. This conservative assumption ensures that, for temperature surface conditions, there are always sufficient non-condensables to maintain a cold trap and prevent excessive water loss; any oxygen accumulation that results is due to other processes.

The model does not include any explicit photochemistry; it tracks fluxes of oxygen into/out of the combined atmosphere-ocean reservoir, and all outgassed reductants are assumed to instantaneously deplete atmospheric oxygen. This simplification is adequate for estimating oxygen accumulation because if oxygen sources exceed oxygen sinks, then oxidant build-up will occur; neither photolysis reactions nor spontaneous reactions can add net reducing power the atmosphere-ocean system. Our model cannot predict low steady state oxygen abundances in predominantly anoxic atmospheres, however; atmospheric O$_2$ is truncated at a lower limit of 0.1 Pa for numerical efficiency. Importantly, our modeling approach is agnostic on the plausibility of the various photochemical scenarios that have been proposed for abiotic O$_2$ atmosphere maintenance by continuous CO$_2$ photodissociation (Gao et al., 2015; Hu et al., 2020). Studies of these scenarios enforce global redox balance at the boundaries of the atmosphere-ocean system and determine whether appreciable atmospheric oxygen exists in the resultant photochemical steady state (e.g., Harman et al., 2015). Here, we are instead using a time-dependent model to investigate whether slight imbalances in atmosphere-ocean boundary fluxes can result in atmospheric oxygen accumulation on long timescales (cf., Luger & Barnes, 2015; Schaefer et al., 2016; Wordsworth et al., 2018).

There are many uncertain parameterizations and parameter values in our model, and so all results are presented as Monte Carlo ensembles that randomly sample a wide range of uncertain parameter values. Parameter ranges and their justifications are described in full in the supporting information (Table S1). We sampled a range of temperature-dependent mantle viscosities, efficiencies of XUV-driven escape, uncertain early sun XUV fluxes, carbon cycle feedbacks, deep hydrological cycle dependencies, and albedo parameterizations. Unknown parameters that are particularly important for oxygen false positives include the dry crustal oxidation efficiency, $f_{\text{dry-oxid}}$, which is the fraction of Fe$^{2+}$ in newly produced crust that is oxidized to Fe$^{3+}$ in the presence of an oxidizing atmosphere via non-aqueous reactions. This parameter is sampled uniformly in log space from $10^{-4}$ to 10% ( Gillmann et al., 2009). Another important parameter is the XUV-driven escape efficiency, $\epsilon_{\text{low-XUV}}$, which is the fraction of stellar XUV energy that drives H-escape. This is sampled uniformly from 0.01 to 0.3, and the portion of energy that goes into escape once the XUV flux exceeds what is required for O-drag is an additional free parameter.

### 2.1. Model Validation

To validate the model, we first show that it can successfully reproduce the atmospheric evolution of Earth and Venus. Venus results are described in detail in supporting information Section C, and here we summarize key results for Earth. Figure 2 shows Monte Carlo model outputs over a range of Earth-like volatile inventories, specifically an initial water content of 1–10 Earth oceans, an initial CO$_2$ content of 20–2,000 bar. Moreover, only initial CO$_2$ inventories less than the initial water inventory by mass are permitted, and an initial (post core-formation) mantle redox state around the Quartz-Fayalite-Magnetite (QFM) buffer is assumed. There is evidence for more reducing Hadean continental crust (Yang et al., 2014), and other terrestrial planets such as Mars likely have reducing mantles (Wadhwa, 2001). While a rapidly oxidized mantle (e.g., Zahnle et al., 2010) is assumed in all nominal calculations, the sensitivity of our results to initial mantle redox is explored in supporting information Section G.
Figure 2a shows the time-evolution of mantle potential temperature and surface temperature, Figure 2b shows the solidification of the magma ocean from core to surface, which takes several Myr, consistent with previous studies. (e) The magma ocean ends when the planet’s interior cools such that heatflow from the interior drops below the runaway greenhouse limit. (d) When this occurs, liquid water oceans condense onto the surface, (f) a temperate carbon cycle commences. (c) There is sometimes a brief spike in atmospheric oxygen following magma ocean solidification due to the persistence of a steam atmosphere and hydrogen escape, (i) but this oxygen is rapidly drawn down by geological sinks. (g) Volatile cycling is controlled by the rate at which fresh crust is produced. Mantle redox evolution is plotted (h) alongside proxy estimates (O’Neill et al., 2018; Trail et al., 2011).

Our modeled early Earth atmosphere-thermal-climate evolution is broadly consistent with semiquantitative reconstructions of Hadean atmospheric evolution (Zahnle et al., 2007, 2010). We also find that in virtually every model run, after 4.5 Gyr of atmospheric evolution, the atmosphere is anoxic (Figure 2c). This is unsurprising. Hydrogen escape during the initial ~ Myr magma ocean does not add free oxygen to the atmosphere but instead oxidizes the interior, as has been described previously (Hamano et al., 2013). In some cases, small amounts of abiotic oxygen are produced in the post magma-ocean steam atmosphere, but...
For Venus, the model can recover current atmospheric conditions assuming the initial water inventory is small, and that crustal sinks of oxygen are efficient (see supporting information Section C). Venusian histories in which the surface was never habitable and in which the surface was habitable for several billion years can both be reconciled with the current atmosphere, which is broadly consistent with previous modeling of Venus’ atmospheric evolution (Chassefière et al., 2012; Kasting & Pollack, 1983; Way et al., 2016).

3. Results

If Earth’s initial volatile inventories are varied, then oxygen-rich atmospheres may be possible. Here, we outline three scenarios whereby an abiotic Earth could have accumulated an oxygen-rich atmosphere after 4.5 Gyr. None of these scenarios guarantee an oxygen-rich atmosphere; instead, oxygenated atmospheres are a possible outcome that is dependent on the efficiency of oxygen crustal sinks and atmospheric escape.

3.1. Scenario 1: High CO$_2$:H$_2$O Initial Inventory Leading to Perpetual Runaway Greenhouse

Figure 3 shows selected model outputs for planets with initial CO$_2$:H$_2$O volatile inventories greater than one by mass and atmospheric O$_2$ > 10$^{17}$ kg at present (P$_{O2}$ > ~0.02 bar). For these planets, the greenhouse warming from a dense CO$_2$ atmosphere ensures that the surface temperature is above the critical point of water; liquid water never condenses on the surface at 1 AU, except briefly, and in small amounts, during this atmospheric oxygen is rapidly overwhelmed by outgassing and other crustal sinks. Subsequent oxygen production via diffusion-limited escape is small, and so there are no further opportunities for abiotic accumulation so long as the planet remains geologically active.

Figure 3. Oxygen false positives from high initial CO$_2$:H$_2$O inventories (Scenario 1). The model is applied to the Earth from magma ocean to present with randomly sampled initial water inventories ranging from ∼0.1 to 10 Earth oceans, and initial CO$_2$ inventories ranging from ∼20 to 2,000 bar (implying CO$_2$:H$_2$O ranging from 0.01 to 100 by mass). Only model outputs with modern day atmospheric oxygen exceeding 10$^{17}$ kg (>∼0.02 bar) are plotted. Subplots are the same as in Figure 2, and shaded regions denote 95% confidence intervals. (a) High atmospheric CO$_2$ ensures the surface temperature always exceeds the critical point of water after the pre-main sequence, and (d) thus permanent liquid water oceans do not condense. The lack of surface water, low volatile content of the mantle, and high surface pressure increasing volatile solubility in partial melts all limits oxygen sinks. (a, i) The largest atmospheric sink is dry crustal oxidation, which diminishes with time as the interior cools. (c) Atmospheric oxygen produced via H escape may start to accumulate after several Gyr of evolution.
These high CO$_2$:H$_2$O perpetual runaway atmospheres have been described previously (Marcq et al., 2017; Salvador et al., 2017). The lack of liquid surface water precludes CO$_2$-draw-down via silicate weathering (Figure 3f). Reactions between supercritical water and silicates will be severely kinetically limited by sluggish solid state diffusion, and are therefore assumed to be negligible (Zolotov et al., 1997). Consequently, a dense CO$_2$ atmosphere and supercritical surface temperature persist indefinitely (Figure 3a), despite the planet residing in the habitable zone. Moreover, there is sufficient steam in the atmosphere to ensure diffusion-limited hydrogen escape provides an appreciable source flux of oxygen (Figure 3i).

Oxygen accumulation also requires limited oxygen sinks, and this may occur on high CO$_2$:H$_2$O worlds for two reasons. First, since permanent oceans do not condense, it is difficult to sequester outgassed volatiles left over from the magma ocean in the interior; hydration reactions and carbonatization reactions do not occur without liquid surface water. Limited mantle regassing after magma ocean outgassing implies low mantle volatile content, which inhibits the capacity of outgassed reductants to draw down crust oxidized. The observed range in carbonaceous chondrite CO$_2$:H$_2$O ratios (purple interval) is shown in (b) as a rough proxy for Earth's initial volatile inventory. The outliers with high oxygen (red box) are Scenario 3 (desertworld) false positives, which are examined in Section 3.3. Anoxic atmospheres truncate at $\sim 10^{-6}$ bar for numerical efficiency (see supporting information); these model runs represent outcomes with essentially no atmospheric oxygen.

However, even with limited outgassing, new crust is still being produced that may be directly oxidized by gaseous O$_2$ (Figure 3g). Figure 4a shows atmospheric oxygen abundances after 4.5 Gyr as a function of dry crustal oxidation efficiency, $f_{dry-oxid}$, for a large number of model runs sampling $10^{20} - 10^{22}$ kg CO$_2$ and $10^{20} - 10^{22}$ kg H$_2$O. Atmospheric oxygen at 4.5 Gyr is plotted as a function of (a) dry crustal oxidation efficiency, and (b) the initial CO$_2$:H$_2$O inventory by mass. Note that Scenario 1 oxygen accumulation (high CO$_2$, perpetual runaway greenhouse atmospheres) requires both an initial CO$_2$:H$_2$O ratio >1 (green box) and for dry crustal oxidation to be relatively inefficient, with <0.1% of Fe$^{2+}$ in newly produced crust oxidized. The observed range in carbonaceous chondrite CO$_2$:H$_2$O ratios (purple interval) is shown in (b) as a rough proxy for Earth's initial volatile inventory. The outliers with high oxygen (red box) are Scenario 3 (desertworld) false positives, which are examined in Section 3.3. Anoxic atmospheres truncate at $\sim 10^{-6}$ bar for numerical efficiency (see supporting information); these model runs represent outcomes with essentially no atmospheric oxygen.

Figure 4. Conditions required for Scenario 1 oxygen false positive. Each dot denotes a single model run, and model runs are shown for uniformly sampled initial volatile abundances: $10^{20} - 10^{22}$ kg CO$_2$ and $10^{20} - 10^{22}$ kg H$_2$O. Atmospheric oxygen at 4.5 Gyr is plotted as a function of (a) dry crustal oxidation efficiency, and (b) the initial CO$_2$:H$_2$O inventory by mass. Note that Scenario 1 oxygen accumulation (high CO$_2$, perpetual runaway greenhouse atmospheres) requires both an initial CO$_2$:H$_2$O ratio >1 (green box) and for dry crustal oxidation to be relatively inefficient, with <0.1% of Fe$^{2+}$ in newly produced crust oxidized. The observed range in carbonaceous chondrite CO$_2$:H$_2$O ratios (purple interval) is shown in (b) as a rough proxy for Earth's initial volatile inventory. The outliers with high oxygen (red box) are Scenario 3 (desertworld) false positives, which are examined in Section 3.3. Anoxic atmospheres truncate at $\sim 10^{-6}$ bar for numerical efficiency (see supporting information); these model runs represent outcomes with essentially no atmospheric oxygen.


3.2. Scenario 2: Waterworlds

Figure 5 shows selected model outputs for planets with H\textsubscript{2}O volatile inventories between 10 and 230 Earth oceans, and initial CO\textsubscript{2} inventories ranging from \(\sim 20\) to 6,000 bar. Only model outputs with modern day atmospheric oxygen exceeding \(10^{17}\) kg (>\(\sim 0.02\) bar) are plotted. Subplots are the same as in Figure 2, and shaded regions denote 95% confidence intervals. (d) The large surface volatile inventory increases the (g) mantle solidus such that melt production and tectonics shut off shortly after formation. (i) This shuts down all oxygen sinks and (c) allows for the gradual accumulation of oxygen via diffusion-limited H escape over several Gyr.
3.3. Scenario 3: Desertworlds

The final scenario whereby abiotic oxygen could accumulate on habitable zone planets around Sun-like stars occurs for planets with extremely small initial volatile inventories (initial water inventory $< -0.3$ Earth oceans). Figure 7 shows selected model outputs representing this desertworld false positive. The required sequence of events is as follows: the low volatile inventory ensures that the magma ocean freezes quickly (typically $\sim 10^7$ years, Figure 7b), even though the planet is still in a runaway greenhouse state due to high heatflow from the interior (Figure 7e). A steam-dominated atmosphere can therefore persist for a few million years, and oxygen may accumulate during this time because there is no surface magma ocean to dissolve the oxygen. Dry crustal oxidation will remove some oxygen during this steam atmosphere phase, but oxidation will be limited by the rate at which oxygen can diffuse into extrusive lava flows (Figures 7i and 8i). When a shallow ocean does eventually condense out as heatflow from the interior drops below the runaway greenhouse limit (Figures 7d and 7e), oxygen may persist for billions of years if oxygen sinks are small (Figure 7c). Outgassing sinks are limited by the low volatile inventory of the planet, but inefficient dry crustal oxidation is also required post-magma ocean for the oxygen to persist for 4.5 Gyr; the efficiency parameter, $f_{\text{dry-oxid}}$, must be $<0.1\%$ (Figure S1). This habitable scenario is qualitatively different to the uninhabitable Scenario 1 false positives: in the former, atmospheric oxygen accumulates early and gradually declines due to crustal sinks, whereas in the latter, oxygen accumulation takes several Gyrs. Desertworld false positives also require efficient XUV-driven hydrodynamic escape ($>10\%$) during the steam atmosphere phase to produce large atmospheric oxygen abundances after the magma ocean has solidified (Figure S1).

4. Discussion

The modeling approach adopted in this paper has several important caveats and limitations. First, we consider the reasons why abiotic oxygen accumulation could be underestimated in our model.

4.1. Assumptions That May Underestimate Abiotic Oxygen Accumulation

As noted above, the model does not track nitrogen fluxes and instead assumes 1 bar $N_2$ partial pressure throughout. This limits oxygen accumulation by providing a non-condensable background gas to throttle hydrogen escape at the cold trap (Kleinböhl et al., 2018; Wordsworth & Pierrehumbert, 2014). Nitrogen atmospheric evolution for terrestrial planets is highly uncertain, and the evolution of Earth’s atmospheric $N_2$ inventory is poorly constrained (Johnson & Goldblatt, 2018; Stüeken et al., 2016). However, for terrestrial planets that form with low nitrogen inventories, or with most of their nitrogen sequestered in the interior (Wordsworth, 2016), then oxygen accumulation on temperate planets could be a more common outcome than our modeling suggests.

Our nominal outgassing model may overestimate fluxes of reduced gases per unit mass partial melt. This is because we do not account for graphite saturation and redox-dependent partitioning of carbon-bearing species between crystalline and melt phases; we instead assume a constant partition coefficient for relating solid mantle $CO_2$ content and total melt plus gas phase concentrations (e.g., Lebrun et al., 2013). Models of Martian outgassing (Grott et al., 2011), and more generalized terrestrial outgassing models (Ortenzi et al., 2020) both show that reducing mantles tend to outgas fewer volatiles by mass than more oxidized mantles for the same amount of crustal production. Our model does, however, account for the greater reducing power of volcanic outgassing on planets with lower mantle oxygen fugacities (Gaillard & Scaillet, 2014; Wogan et al., 2020). Consequently, our conservative approach maximizes fluxes of outgassed reductants, and may...
underestimate oxygen accumulation. Sensitivity tests which consider more reducing mantles and graphite saturated melts are described below.

Finally, our model may underestimate the duration of steam atmospheres following magma ocean solidification, and therefore underestimate oxygen accumulation prior to ocean condensation. Based on the time required to precipitate an Earth ocean and the apparent absence of stable climate states at the runaway greenhouse limit (Figure S4), it is typically assumed that the time required to transition from steam atmosphere to surface water ocean is \( \sim 10^3 \) years (Abe, 1993; Zahnle et al., 2007). In our model, the atmosphere-ocean system is assumed to be in radiative equilibrium with a negligible heat capacity (Lebrun et al., 2013). However, for waterworlds with hundreds of Earth oceans, the time required for a steam atmosphere to condense is long enough for abiotic oxygen accumulation to be significant. Moreover, stable climate states may exist in between a molten surface and a temperate surface ocean, especially if clouds—which we ignore—are included in radiative transfer calculations (Marcq et al., 2017 their Figure 6). Finally, in our model, oxygen is partitioned between the magma ocean and the atmosphere assuming chemical equilibrium. This is a reasonable assumption for high temperature/low-viscosity magma oceans with very short mixing times, but as the surface temperature approaches the solidus, the “magma ocean mush” will behave more like a solid (Lebrun et al., 2013; Salvador et al., 2017), and oxygen produced during the final stages of the magma ocean may not be efficiently sequestered in the melt.

### 4.2. Assumptions That May Overestimate Abiotic Oxygen Accumulation

Next, we consider model assumptions that might cause abiotic oxygen accumulation to be overestimated. The omission of a nitrogen cycle means the model ignores the removal of atmospheric oxygen in the ocean as dissolved nitrate. Nitrate formation is thermodynamically favorable at temperate surface conditions

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**Figure 7.** Oxygen false positives on desertworlds (Scenario 3). The model is applied to the Earth from magma ocean to present with randomly sampled initial water inventories ranging from \( \sim 0.05 \) to 0.35 Earth oceans, and initial CO\(_2\) inventories ranging from \( \sim 6 \) to 60 bar. Only model outputs with modern day atmospheric oxygen exceeding \( 10^{17} \) kg (\( > \sim 0.02 \) bar) are plotted. Subplots are the same as in Figure 2, and shaded regions denote 95% confidence intervals. (b) The low initial volatile inventory ensures the magma ocean solidifies before the runaway greenhouse is over, (c) allowing for significant oxygen accumulation in the \( \sim \)Myr steam atmosphere that (e) persists until the heatflow from the interior drops below the runaway greenhouse limit. If subsequent oxygen sinks are low, then the few bar oxygen that accumulate early on may persist for billions of years.
(Krissansen-Totton et al., 2016), and the reaction may occur via lightning. Indeed NO-formation via lightning has previously been established as an important mechanism for removing photochemically produced atmospheric oxygen and inhibiting oxygen false positives (Harman et al., 2018). However, nitrogen fixation via lighting is unlikely to prevent abiogenic oxygen accumulation on waterworlds. In the absence of any other sources or sinks nitrate-formation via lightning could draw down the modern Earth's atmospheric oxygen reservoir in 20–200 Myr (Krissansen-Totton et al., 2016) and could therefore remove 1 bar of nitrogen every 0.17–1.7 Gyr. Once most atmospheric nitrogen is converted to nitrate in the ocean, oxygen may accumulate rapidly due to the lack of a non-condensible cold trap (Wordsworth & Pierrehumbert, 2014), and nitrogen will not be significantly replenished by outgassing on waterworlds due to the pressure overburden effect precluding new crustal production. Waterworlds with large initial nitrogen atmospheric inventories (10 s of bar) could avoid abiogenic oxygen accumulation if oxygen drawdown via lightning exceeds oxygen production via diffusion-limited hydrogen escape. However, it is debated whether nitrate is the kinetically stable form of nitrogen on habitable worlds (Hu & Diaz, 2019; Ranjan et al., 2019; Wong et al., 2017). Efficient conversion of nitrate back to molecular nitrogen via abiogenic chemodenitrification might return oxygen to the atmosphere and prevent its accumulation in the ocean, regardless of the initial N2 volatile inventory. In summary, it is unlikely that nitrate sinks will always preclude abiogenic oxygen accumulation on waterworlds, but better constraints on aqueous nitrogen chemistry would improve this assessment.

Another important limitation of our model is that it ignores infrared cooling of the upper atmosphere and the throttling of hydrogen escape that results from a cool stratosphere. In our nominal model, an isothermal 210 K stratosphere is assumed. However, CO2-rich atmospheres may efficiently radiate in the IR, cooling the stratosphere and enhancing the water cold trap (Wordsworth & Pierrehumbert, 2013). To test the sensitivity of our results to stratospheric temperature, we repeated our calculations and introduced an additional stratospheric temperature variable, which was randomly sampled from 150 to 250 K. The full results of these calculations are shown in Figure S13. To summarize, neglecting the stratospheric radiation budget does not affect the viability of Scenarios 2 (waterworlds) or 3 (desertworlds) because, in the former, the oxygen source is diffusion-limited escape through a N2-dominated atmosphere at modern Earth-like rates, whereas for the later, oxygen accumulation occurs predominantly during the early, steam-dominated atmosphere, and so stratospheric temperature does not have a strong influence on escape rates (Figures S13b and S13c). However, oxygen accumulation via Scenario 1 (high CO2; H2O perpetual runaway greenhouse) is unlikely if the stratosphere is cooler than 200 K (Figure S13a). Photochemically produced ozone or hazes may offset the cooling effect of CO2, and so a full assessment of Scenario 1 requires more detailed radiative-photochemical modeling.

One additional caveat is that our model assumes an anhydrous solidus. This simplification is probably reasonable for post-magma ocean mantle conditions (Kite & Ford, 2018), but it is possible to imagine a scenario whereby waterworld mantles become increasingly hydrated via subduction after the magma ocean phase, and that this hydration offsets the pressure overburden effect to maintain geologic activity, and therefore oxygen sinks, for much longer than our nominal model suggests. To test this possibility, we conducted a sensitivity test where we accounted for mantle hydration decreasing the solidus (Katz et al., 2003). The results of this sensitivity test are discussed in detail in supporting information Section D, but in summary, mantle hydration does not have a large effect on oxygen accumulation on waterworlds; it merely shifts the ocean mass threshold for oxygen accumulation.

Sensitivity tests were also conducted to assess whether the delivery of reducing material such as metallic iron and FeO via impacts (Zahnle et al., 2020), a more reducing initial mantle, or larger planet-star separations could inhibit abiogenic oxygen accumulation. These sensitivity test results are described in full in supporting information Sections E, G, and H, respectively. In summary, we find that high impactor fluxes could preclude desertworld oxygen accumulation assuming all impactor material is completely oxidized (supporting information Section E). Impactors may thus prevent Scenario 3 (desertworld) false positives in some cases, although this is not guaranteed because it is possible to imagine planetary formation pathways with smaller impactor fluxes and/or where the majority of impactor material is buried or lost to space as suggested by some impact simulations (Marchi et al., 2018). Scenarios 1 (high CO2; H2O perpetual runaway greenhouse) and 2 (waterworlds) are viable under a more reducing (iron-wüstite buffer) initial mantle (supporting information Section G). This counterintuitive result occurs because, even though degassed
volatiles are likely to be more reducing, total volatile concentrations in the melt phase are typically lower due to graphite saturation (Grott et al., 2011; Ortenzi et al., 2020). Moreover, crustal sinks are precluded by high overburden pressure, regardless of the redox state of the crustal material. Although Scenario 3 (desertworlds) is seemingly not excluded by a more reducing mantle, evaluating this would require more complete radiative transfer and photochemical modeling of CO-H2 dominated atmospheres. Finally, when nominal calculations are repeated at 1.3 AU, both Scenario 2 and Scenario 3 oxygen false positives still occur frequently (supporting information Section H). Scenario 1 false positives do not occur at large planet-star separations because a high CO2:H2O atmosphere cannot maintain a perpetual runaway greenhouse state after magma ocean solidification.

For Scenario 1 and 3 to be viable, the dry crustal oxidation parameter, \( f_{\text{dry-oxid}} \), must be relatively small (<0.1%). This contrasts with Venus where \( f_{\text{dry-oxid}} \) probably needs to exceed >0.1% to remove virtually all \( O_2 \) from the atmosphere (see Venus validation in supporting information). This parameter is challenging to definitively constrain because it represents a range of physical processes including the diffusion of oxygen into extrusive lava flows (Gillmann et al., 2009), direct oxidation of small grain erosion products (Arvidson et al., 1992), and various other gas-solid redox reactions (Zolotov, 2019). Even if the oxidation of fresh crust is typically efficient, low dry crustal oxidation efficiencies cannot be ruled out because tectonic regimes where most magmatic activity is intrusive and isolated from the atmosphere are possible. In any case, the uncertainties in crustal oxidation processes highlights need for future missions to Venus to better constrain its redox evolution. Mars’ crust is more oxidized than its upper mantle, but this oxidation cannot necessarily be used to constrain dry crustal oxidation efficiency since it might be attributable to early hydrous alteration (Herd et al., 2002; Wadhwa, 2001). The presence of gray, reduced sediments mere centimeters below more oxidized Martian regolith, as revealed by Curiosity, argues against efficient post depositional gas-solid oxidation under oxic conditions (Ming et al., 2014).

Finally, we note that in some cases abiotic oxygen accumulation is contingent on highly uncertain atmospheric escape physics. This uncertainty does not affect the viability of the waterworld (Scenario 2) false positives because the required H escape flux is comparable to the modern Earth’s diffusion-limited escape flux (Catling & Kasting, 2017, p. 148). Oxygen accumulation only occurs in this case because of the pressure overburden suppression of oxygen sinks. Scenario 1 (high CO2:H2O perpetual runaway greenhouse) is similarly unaffected. However, for Scenario 3 (desertworlds), oxygen accumulation only occurs because of efficient XUV-driven escape of hydrogen, \( \epsilon_{\text{XUV}} > 0.1 \). If H-escape is photochemically limited (e.g., Wordsworth et al., 2018) then oxygen accumulation may be limited by the photochemical dissociation of water by UV photons, and the rate and which \( H_2O \) recombination reactions occur. However, the oxygen source fluxes required in our desertworld scenario (~500 Tmol \( O_2 \)/yr, Figure 7i) are comparable to the water loss rates inferred for a steam-only early Earth atmosphere calculation using a photochemical model (Wordsworth et al., 2018, their Figure 9). Future work ought to couple the geochemical evolution model here to a photochemical model that includes C-bearing species to better assess the potential for oxygen accumulation on desertworlds. Finally, it is possible that non-thermal O loss (Airapatian et al., 2017) or photochemically modulated stoichiometric escape of \( H \) and \( O \) (McElroy, 1972) could lessen oxygen accumulation. Upcoming JWST observations of highly irradiated terrestrial planets may constrain escape processes and improve predictions of oxygen accumulation on more temperate planets.

Sulfur outgassing and burial may have played an important role in the oxygenation of Earth’s atmosphere (Gaillard et al., 2011; Olson et al., 2019). Sulfur species are ignored in our nominal model because their bulk abundances are probably too small to qualitatively change our oxygenation scenarios (Wordsworth et al., 2018). Supporting information Section I explores the consequences of adding reduced sulfur species to our outgassing model, and confirms that, for Earth-like sulfur mantle abundances, total oxygen sinks are comparable to when sulfur is neglected. With that said, mantle sulfur abundances are contingent on formation processes (e.g., Grewal et al., 2019) and could be highly variable. Incorporating a complete sulfur cycle into a redox evolution model to investigate the sensitivity of oxygenation to initial sulfur content is an opportunity for future research. Note, however, that mantle sulfur abundances are irrelevant for waterworlds (Scenario 2) where all crustal production is suppressed.

In summary, there are several unknowns that preclude definitive predictions of how frequently the three scenarios outlined in this study might occur, but none can be ruled-out with current knowledge.
4.3. Implications for Future Observations

How might future observations discriminate between the three abiotic oxygen scenarios described above and oxygen produced by a biosphere? In principle, high CO₂:H₂O atmospheres should be possible to diagnose via direct imaging spectral observations because they are not habitable. A clear atmosphere is likely since the coexistence of atmospheric H₂O, O₂, O₃, and abundant OH radicals may preclude the accumulation of photochemical hazes. Strong CO₂ absorption features ought to be visible, as should pressure-sensitive CIA features from the high pressures; more detailed photochemical and spectroscopic simulations will be required to determine the best false positive discriminants for these worlds.

Because waterworlds and desertworlds are habitable, they may be more challenging to discriminate from inhabited terrestrial planets. Crucially, waterworld false positives would be ruled out by a detection of subaerial land because, for Earth-like gravity, the presence of emerged continents limits the maximum ocean depth to around 10 km (Cowan & Abbot, 2014), or equivalently, a few Earth oceans by mass. This limit arises because silicates cannot support their own weight with greater topography. Consequently, the detection of an ocean-continent dichotomy using time-resolved photometric mapping (Cowan et al., 2009; Farr et al., 2018; Fujii et al., 2010; Kawahara & Fujii, 2010; Lustig-Yaeger et al., 2018) could rule out a waterworld false positive, assuming alternative explanations for dichotomies in surface maps could be excluded. This highlights the need for large aperture direct imaging missions to ensure sufficient time-resolution to map the surface over a planet’s rotation. Alternatively, independent mass and radius constraints from radial velocity observations and thermal infrared direct imaging (Quanz et al., 2019), respectively, could also help rule out large (few wt.%) water inventories based on bulk density.

Desertworlds are likely the most challenging scenario to disambiguate from biological oxygen. Time-resolved photometric surface maps and/or the lack of ocean glint could help evaluate the surface water inventory and might be suggestive of a small water inventory (Lustig-Yaeger et al., 2018; Robinson et al., 2010). There are potentially other diagnostic spectral signatures of desertworlds such as spatial variation in atmospheric water vapor and photochemistry that could be tested using general circulation models and photochemical models. The presence of long-lived sulfuric acid hazes (Loftus et al., 2019) has been proposed as putting an upper bound on surface water abundances, but the desertworlds considered here likely have larger surface water inventories than this threshold.

Broadly speaking, the scenarios outlined in this study emphasize that no single observation, including oxygen detection on habitable zone planets around sun-like stars, will be uniquely diagnostic of life. It will be necessary to design future telescopes that are capable of both constraining the full planetary/stellar context and identifying multiple lines of evidence for life (Catling et al., 2018; Walker et al., 2018). For example, oxygen detection on an ostensibly habitable terrestrial planet would be persuasive if accompanied by surface biosignature detections (Schwieterman et al., 2018), temporal biosignatures (Olson et al., 2018), or co-existing reducing gases in atmospheric disequilibrium (Krissansen-Totton et al., 2016). The coexistence of oxygen and methane remains an excellent biosignature and would not be expected for any of the oxygen false-positive scenarios described above. Indeed, it is difficult to produce large methane abundances in habitable planet atmospheres without life, even in anoxic atmospheres (Krissansen-Totton, Olson, et al., 2018; Wogan et al., 2020). It should also be noted that the scenarios in this study were illustrated for habitable zone planets around sun-like stars, but they may also be applicable to habitable zone planets around M-dwarfs.

5. Conclusions

The redox evolution of habitable zone terrestrial planets is strongly dependent on initial volatile inventories and the efficiency of crustal sinks. Uninhabited, Earth-sized planets within the habitable zone of G-type stars are very unlikely to accumulate abiotic oxygen if their initial volatile inventories are Earth-like. However, if initial volatile inventories differ dramatically from that of the Earth, then non-biological oxygen accumulation is possible, even when atmospheric noncondensable inventories are large. This may occur when either (i) the initial CO₂:H₂O ratio exceeds one, which suppresses oxygen sinks due to the low mantle volatile content and because surface conditions are too hot for aqueous reactions, or (ii) the initial H₂O inventory is very large, thereby halting crustal production after a few billion years and shutting off all
oxygen sinks, or (iii) the planet is very volatile-poor, in which case oxygen may accumulate during the steam atmosphere that persists after magma ocean solidification. Inefficient dry crustal oxidation is required for scenarios (i) and (iii) to yield large oxygen abundances, and scenario (i) is sensitive to stratospheric temperature. Fortunately, observational discriminants exist for all three of these scenarios; scenario (i) planets are uninhabitable, whereas the ability to constrain surface water inventories using time-resolved photometry would be useful for ruling out scenarios (ii) and (iii). More generally, the possible existence of these oxygen false positive scenarios highlights the need for a systems approach to biosignature assessment where biogenicity is judged not by the presence or absence of a single biosignature gas, but by multiple lines of evidence from both spectrally resolved and temporally resolved observations.

Conflict of Interest
The authors declare no conflicts of interest relevant to this study.

Data Availability Statement
The Python code for our model is open source, https://doi.org/10.5281/zenodo.4539040.

References
Abe, Y. (1993). Physical state of the very early Earth. Lithos, 30(3–4), 223–235. https://doi.org/10.1016/0024-4937(93)90037-d
Airapetian, V. S., Glocer, A., Khazanov, G. V., Loyd, R. O. P., France, K., Sofka, J., et al. (2017). How hospitable are space weather affected habitable zones? The role of ion escape. The Astrophysical Journal Letters, 836(1, L3).
Arney, G., Batalha, N., Britt, A. V., Cowan, N., Domagal, G., Shown, D., et al. (2019). The Sun-like Stars opportunity. Bulletin of the American Astronomical Society, 51(3), 91.
Asad, R. E., Gleeley, R., Malin, M. C., Saunders, R. S., Izenberg, N., Plaut, J. J., et al. (1992). Surface modification of Venus as inferred from Magellan observations of plains. Journal of Geophysical Research, 97(E8), 13303–13317. https://doi.org/10.1029/92je01384
Baraffe, I., Chabrier, G., Allard, F., & Hauschildt, P. H. (1998). Evolutionary models for solar metallicity low-mass stars: Mass-magnitude relationships and color-magnitude diagrams. Astronomy and Astrophysics, 337, 403–412.
Baraffe, I., Chabrier, G., Allard, F., & Hauschildt, P. H. (2003). Evolutionary models for low-mass stars and brown dwarfs: Uncertainties and limits at very young ages. Astronomy and Astrophysics, 182(2), 563–572. https://doi.org/10.1051/0004-6361:2001638
Barth, P., Carone, L., Barnes, R., Noack, L., Mollière, P., & Henning, T. (2020). Magma ocean evolution of the TRAPPIST-1 planets. arXiv preprint. arXiv:2008.09599.
Benneke, B., Wong, I., Plaulet, C., Knutson, H. A., Lothringer, J., Morley, C. V., et al. (2019). Water vapor and clouds on the habitable-zone super-Neptune exoplanet K2-18b. The Astrophysical Journal Letters, 887(1), L14. https://doi.org/10.3847/2041-8213/ab59dc
Buick, R. (2008). When did oxygenic photosynthesis evolve? Philosophical Transactions of the Royal Society of London. Series B Biological Sciences, 363(1504), 2731–2743. https://doi.org/10.1098/rstb.2008.0084
Catling, D. C., & Kasting, J. F. (2017). Earth-like oceans and biosignatures. The mosaics of life. Cambridge University Press.
Catling, D. C., Krissansen-Totton, J., Kiang, N. Y., Crisp, D., Robinson, T. D., DasSarma, S., et al. (2018). Exoplanet biosignatures: A framework for their assessment. Astrobiology, 18(6), 709–738. https://doi.org/10.1089/ast.2017.17377
Chassefière, E., Wieler, R., Marty, B., & Leblanc, F. (2012). The evolution of Venus: Present state of knowledge and future exploration. Planetary and Space Science, 63–64, 15–23. https://doi.org/10.1016/j.pss.2011.04.007
Cowan, N. B., & Abbot, D. S. (2014). Water cycling between ocean and mantle: Super-Earths need not be waterworlds. The Astrophysical Journal, 781(1), 27. https://doi.org/10.1088/0004-637x/781/1/27
Cowen, N. B., Agol, E., Meadows, V. S., Robinson, T., Livengood, T. A., Deming, D., et al. (2009). Alien maps of an ocean-bearing world. The Astrophysical Journal, 700(2), 915. https://doi.org/10.1088/0004-637x/700/2/915
Diamond-Lowe, H., Charbonneau, D., Malik, M., Kempton, E. M.-R., & Beletsky, Y. (2020). Optical transmission spectroscopy of the terrestrial exoplanet LHS 3844b from 13 ground-based transit observations. arXiv preprint. arXiv:2008.05444.
Domagal-Goldman, S. D., Segura, A., Claire, M. W., Robinson, T. D., & Meadows, V. S. (2014). Abiotic ozone and oxygen in atmospheres similar to prebiotic Earth. The Astrophysical Journal, 782(2), 90. https://doi.org/10.1088/0004-637x/782/2/90
Dong, C., Jin, M., Lingam, M., Airapetian, V. S., Ma, Y., & van der Holst, B. (2018). Atmospheric escape from the TRAPPIST-1 planets and implications for habitability. Proceedings of the National Academy of Sciences of the United States of America, 115(2), 260–265. https://doi.org/10.1073/pnas.1708010115
Farr, B., Farr, W. M., Cowan, N. B., Haggard, H. M., & Robinson, T. (2018). Exocartographer: A Bayesian framework for mapping exoplanets in reflected light. The Astronomical Journal, 156(4), 146. https://doi.org/10.3847/1538-3881/aad775
Faucher-Giguère, C. A., Villanueva, G. L., Schwieterman, E. W., Turbet, M. J., Arney, G., Pichierri, D., et al. (2020). Sensitive probing of exoplanetary atmospheres via mid-infrared collisional absorption. Nature Astronomy, 4, 372–376.
Feng, Y. K., Robinson, T. D., Fortney, J. J., Luhu, E. R., Marley, M. S., Lewis, N. K., et al. (2018). Characterizing Earth analogs in reflected light: Atmospheric retrieval studies for future space telescopes. The Astronomical Journal, 155(5), 200. https://doi.org/10.3847/1538-3881/aab96c
Fischer, D., Bradley, P., Bean, J., Calzetti, D., Dawson, R., Dressing, C., et al. (2019). The LUVOIR Mission Concept study final report. arXiv preprint. arXiv:1912.06219.
Pufii, Y., Kawahara, H., Suto, Y., Taruya, A., Fukuda, S., Nakajima, T., & Turner, E. L. (2010). Colors of a second Earth: Estimating the fractional areas of ocean, land, and vegetation of Earth-like exoplanets. The Astrophysical Journal, 719(2), 866. https://doi.org/10.1088/0004-637x/719/2/866
Gaillard, F., & Scallet, B. (2014). A theoretical framework for volcanic degassing chemistry in a comparative planetology perspective and implications for planetary atmospheres. *Earth and Planetary Science Letters*, 403, 307–316. https://doi.org/10.1016/j.epsl.2014.07.009

Gaillard, F., Scallet, B., & Arndt, N. T. (2011). Atmospheric oxygenation caused by a change in volcanic degassing pressure. *Nature*, 478(7368), 229. https://doi.org/10.1038/nature10460

Gao, P., Hu, R., Robinson, T. D., Li, C., & Yung, Y. L. (2015). Stability of CO2 atmospheres on desiccated M Dwarf Exoplanets. *The Astrophysical Journal*, 806, 249–261. https://doi.org/10.1088/0004-637x/806/2/249

Gaudi, B. S., Seager, S., Mennessou, K., Kessling, A., Warfield, K., Cahoy, K., et al. (2020). The Habitable Exoplanet Observatory (HabEx) Mission Concept study final report. arXiv preprint. arXiv:2001.06683.

Gillmann, C., Chassefière, E., & Logonné, P. (2009). A consistent picture of early hydrodynamic escape of Venus atmosphere explaining present Ne and Ar isotopic ratios and low oxygen atmospheric content. *Earth and Planetary Science Letters*, 286(3–4), 503–513. https://doi.org/10.1016/j.epsl.2009.07.016

Grewal, D. S., Dasgupta, R., Sun, C., Tuzzo, K., & Costin, G. (2019). Delivery of carbon, nitrogen, and sulfur to the silicate Earth by a giant impact. *Science Advances*, 5(1), eaau3669. https://doi.org/10.1126/sciadv.aau3669

Grimm, S. L., Demory, B.-O., Gillon, M., Dorn, C., Agol, E., Burdanov, A., et al. (2018). The nature of the TRAPPIST-1 exoplanets. *Astronomy & Astrophysics*, 613, A68. https://doi.org/10.1051/0004-6361/201732233

Grott, M., Morschhauser, A., Breuer, D., & Hauber, E. (2011). Volcanic outgassing of CO2 and H2O on Mars. *Earth and Planetary Science Letters*, 308(3–4), 391–400. https://doi.org/10.1016/j.epsl.2011.06.014

Hamano, K., Abe, Y., & Genda, H. (2013). Emergence of two types of terrestrial planet on solidification of magma ocean. *Nature*, 497(7445), 607–610. https://doi.org/10.1038/nature12163

Harman, C. E., Felton, R., Hu, R., Domagal-Goldman, S. D., Segura, A., Tian, F., & Kasting, J. F. (2018). Abiotic O2 levels on planets around F, G, K, and M stars: Effects of lightning-produced catalysts in eliminating oxygen false positives. *The Astrophysical Journal*, 866(1), 56.

Harman, C. E., Schwieterman, E. W., Schottelkotte, J. C., & Kasting, J. F. (2015). Abiotic O2 levels on planets around F, G, K, and M Stars: Possible false positives for life? *The Astrophysical Journal*, 812(2), 137. https://doi.org/10.1088/0004-637x/812/2/137

Herd, C. D. K., Borg, L. E., Jones, J. H., & Papike, J. J. (2002). Oxygen fugacity and geochemical variations in the Martian basalts: Implications for Martian basin petrogenesis and the oxidation state of the upper mantle of Mars. *Geochimica et Cosmochimica Acta*, 66(11), 2025–2036. https://doi.org/10.1016/s0016-7037(02)00986-3

Holland, H. D. (2006). The oxygenation of the atmosphere and oceans. *Philosophical Transactions of the Royal Society of London. Series B Biological Sciences*, 361(1470), 903–915. https://doi.org/10.1098/rstb.2006.1838

Hu, R., & Diaz, H. D. (2019). Stability of nitrogen in planetary atmospheres in contact with liquid water. *The Astrophysical Journal*, 886(2), 126. https://doi.org/10.3847/1538-4357/abcc6a

Hu, R., Peterson, L., & Wolf, E. T. (2020). O2- and CO-rich atmospheres for potentially habitable environments on TRAPPIST-1 Planets. *The Astrophysical Journal*, 898(2), 122. https://doi.org/10.3847/1538-4357/ab507b

Johnson, B. W., & Goldblatt, C. (2018). EarthIn: A new Earth system nitrogen model. *Geochimica, Geophysics, Geosystems*, 19(8), 2516–2542. https://doi.org/10.1029/2017gc007392

Kane, S. R., Roetenbacher, R. M., Unterborn, C. T., Foley, B. J., & Hill, M. L. (2020). A volatile-poor formation of LHS 3844b based on its lack of significant atmosphere. *The Planetary Science Journal*, 1(2), 36. https://doi.org/10.3847/psj/abaab5

Kasting, J. F. (1988). Runaway and moist greenhouse atmospheres and the evolution of Earth and Venus. *Icarus*, 74(3), 472–494. https://doi.org/10.1016/0019-1035(88)90116-9

Kasting, J. F., & Pollack, J. B. (1983). Loss of water from Venus. I. Hydrodynamic escape of hydrogen. *Icarus*, 53(3), 479–508. https://doi.org/10.1016/0019-1035(83)90212-9

Katz, R. F., Spiegelman, M., & Langmuir, C. H. (2003). A new parameterization of hydrous mantle melting. *Geochimica, Geophysics, Geosystems*, 4, 1073. https://doi.org/10.1029/2002GC000433

Kawahara, H., & Fujii, Y. (2010). Global mapping of Earth-like exoplanets from scattered light curves. *The Astrophysical Journal*, 720(2), 1333. https://doi.org/10.1088/0004-637x/720/2/1333

Kite, E. S., & Ford, E. B. (2018). Habitablety of exoplanet waterworlds. *The Astrophysical Journal*, 864(1), 75. https://doi.org/10.3847/1538-4357/aade60

Kleinböhl, A., Willacy, K., Friedson, A. J., Chen, P., & Swain, M. R. (2018). Buildup of abiotic oxygen and ozone in moist atmospheres of temperate terrestrial exoplanets and its impact on the spectral fingerprint in transit observations. *The Astrophysical Journal*, 862(2), 92. https://doi.org/10.3847/1538-4357/aaca36

Kreidberg, L., Koll, D. B., Morley, C., Hu, R., Schaerfer, L., Deming, D., et al. (2019). Absence of a thick atmosphere on the terrestrial exoplanet HD 209458b. *The Astrophysical Journal*, 887(2), 348. https://doi.org/10.3847/1538-4357/ab4f4a

Lebrun, T., Massol, H., Chassefière, E., Davaille, A., Marocq, E., Sarda, P., et al. (2013). Thermal evolution of an early magma ocean in interaction with the atmosphere. *Journal of Geophysical Research: Planets*, 118(6), 1155–1176. https://doi.org/10.1002/jgre.20068

Leung, M., Meadows, V. S., & Lustig-Yaeger, J. (2020). High-resolution spectral discriminants of ocean loss for M-dwarf terrestrial exoplanets. *The Astronomical Journal*, 160(1), 11. https://doi.org/10.3847/1538-3881/ab9012

Lofthus, K., Wordsworth, R. D., & Morley, C. V. (2019). Sulfate aerosol hazes and SO2 gas as constraints on Rocky Exoplanets' surface liquid water. *The Astrophysical Journal*, 887(2), 331. https://doi.org/10.3847/1538-4357/ab58cc

Luger, R., & Barnes, R. (2015). Extreme water loss and abiotic O2 buildup on planets throughout the habitable zones of M dwarfs. *Astrobiology*, 15(2), 119–143. https://doi.org/10.1089/ast.2014.1231

Lustig-Yaeger, J., Meadows, V. S., & Lincolnski, A. P. (2019). The detectability and characterization of the TRAPPIST-1 exoplanet atmospheres with JWST. *The Astronomical Journal*, 158(1), 27. https://doi.org/10.3847/1538-3881/aab21e

Lustig-Yaeger, J., Meadows, V. S., Tovar-Mendoza, G., Schwieterman, E. W., Fuji, Y., Luger, R., & Robinson, T. D. (2018). Detecting ocean glint on exoplanets using multiphase mapping. *The Astronomical Journal*, 156(6), 301. https://doi.org/10.3847/1538-3881/a5e3a
Lyons, T. W., Reinhard, C. T., & Planavsky, N. J. (2014). The rise of oxygen in Earth's early ocean and atmosphere. *Nature*, 506(7488), 307–315. https://doi.org/10.1038/nature13068

Marchi, S., Canup, R. M., & Walker, R. J. (2018). Heterogeneous delivery of silicate and metal to the Earth by large planetesimals. *Nature Geoscience*, 11(1), 77–81. https://doi.org/10.1038/s41561-017-0022-3

Marcq, E., Salvador, A., Massol, H., & Davaille, A. (2017). Thermal radiation of magma ocean planets using a 1-D radiative-convective model of H₂O-CO₂ atmospheres. *Journal of Geophysical Research: Planets*, 122(7), 1539–1553. https://doi.org/10.1002/2016je005224

McElroy, M. B. (1972). Mars: An evolving atmosphere. *Science*, 175(4020), 443–445. https://doi.org/10.1126/science.175.4020.443

Meadows, V. S., Reinhard, C. T., Arney, G. N., Parenteau, M. N., Schwiterman, E. W., Domagal-Goldman, S. D., et al. (2018). Exoplanet biosignatures: Understanding oxygen as a biosignature in the context of its environment. *Astrobiology*, 18(6), 630–662. https://doi.org/10.1089/ast.2017.1727

Meixner, M., Cooray, A., Leisawitz, D., Staguhn, J., Armus, L., Batterby, C., et al. (2019). *Origins space telescope mission concept study report*. arXiv preprint. arXiv:1912.06213.

Ming, D. W., Archer, P. D., Gladin, D. P., Eigenbrode, J. L., Franz, H. B., Setter, B., et al. (2014). Volatile and organic compositions of sedimentary rocks in Yellowknife Bay, Gale Crater, Mars. *Science*, 343(6169), 1245267.

Mulikidjianian, A. Y., Koonin, E. V., Makarova, K. S., Mechkov, S. L., Sorokin, A., Wolf, Y. I., et al. (2006). The cyanobacterial genome core and the origin of photosynthesis. *Proceedings of the National Academy of Sciences of the United States of America*, 103(35), 13126–13131. https://doi.org/10.1073/pnas.0605790103

National Academies of Sciences, Engineering, & Medicine. (2018). *Exoplanet science strategy*. National Academies Press.

Nicholls, R. J., McCay, T. J., Clark, P. E., Murphy, M. E., Trombka, J. I., & Jarosewich, E. (2004). Bulk elemental compositions of meteorites: A guide for interpreting remote-sensing geochemical measurements of planets and asteroids. *Antarctic Meteorite Research*. Noack, L., Höning, D., Rivoldini, A., Heistracher, C., Zimov, N., Journaux, B., et al. (2016). Water-rich planets: How habitable is a water layer deeper than on Earth? *Icarus*, 277, 215–236. https://doi.org/10.1016/j.icarus.2016.05.009

Ortenzi, G., Noack, L., Sohl, F., Guimond, C. M., Grenfell, J. L., Dorn, C., et al. (2020). Mantle redox state drives outgassing chemistry and atmospheric composition of rocky planets. *Scientific Reports*, 10(1), 1–14.

Pluriel, W., Marcq, E., & Turbet, M. (2019). Modeling the albedo of Earth-like magma ocean planets with H₂O-CO₂ atmospheres. *Icarus*, 317, 583–590. https://doi.org/10.1016/j.icarus.2018.08.023

Quanz, S. P., Absil, O., Angerhausen, D., Benz, W., Bonfils, X., Berger, J.-P., et al. (2019). *Atmospheric characterization of terrestrial exoplanets in the mid-infrared: Biosignatures, habitability & diversity*. arXiv preprint. arXiv:1908.03116.

Rajanian, S., Schwiterman, E. W., Harman, C., Fateev, A., Sousa-Silva, C., Seager, S., & Hu, R. (2020). Photochemistry of anoxic arboitic Habitable planet atmospheres: Impact of new H₂O cross sections. *The Astrophysical Journal*, 896(2), 148. https://doi.org/10.3847/1538-4357/ab9363

Rajanian, S., Todd, Z. R., Rimmer, P. B., Sasselov, D. D., & Babinb, A. R. (2019). Nitrogen oxide concentrations in natural waters on early Earth. *Geochemistry, Geophysics, Geosystems*, 20(4), 2021–2039. https://doi.org/10.1002/2018gc008082

Raymond, S. N., Kokubo, E., Morbidelli, A., Morishima, R., & Walsh, K. J. (2014). Terrestrial planet formation at home and abroad. In H. Beuther R, Klessen C. Dullemond & Th. Henning (Eds.), *Protostars and Planets VI*. University of Arizona Press.

Raymond, S. N., Quinn, T., & Lunine, J. I. (2004). Making other earths: Dynamical simulations of terrestrial planet formation and water delivery. *Icarus*, 168(1), 1–17. https://doi.org/10.1016/j.icarus.2003.11.019

Ribas, R. K. (2015). Modeling siderophile elements during core formation and accretion, and the role of the deep mantle and volatiles. *American Mineralogist*, 100(5–6), 1098–1109. https://doi.org/10.2138/am-2015-5052

Robinson, T. D., Meadows, V. S., & Crisp, D. (2010). Detecting oceans on extrasolar planets using the glint effect. *The Astrophysical Journal Letters*, 721(1), L67. https://doi.org/10.1088/2041-8205/721/1/L67

Rodler, F., & López-Morales, M. (2014). Feasibility studies for the detection of O₂ in an Earth-like exoplanet. *The Astrophysical Journal*, 781(1), 54. https://doi.org/10.1088/0004-637X/781/1/54

Salvador, A., Massol, H., Davaille, A., Marcq, E., Sarda, P., & Chassefière, E. (2017). The relative influence of H₂O and CO₂ on the primitive surface conditions and evolution of rocky planets. *Journal of Geophysical Research: Planets*, 122(7), 1458–1486. https://doi.org/10.1002/2017je005286

Schafer, L., & Sasselov, D. (2015). The persistence of oceans on Earth-like planets: Insights from the deep-water cycle. *The Astrophysical Journal*, 801(1), 40. https://doi.org/10.1088/0004-637x/801/1/40

Schafer, L., Wordsworth, R. D., Berta-Thompson, Z., & Sasselov, D. (2016). Predictions of the atmospheric composition of GI 1132b. *The Astrophysical Journal*, 829(2), 63. https://doi.org/10.3847/0004-637x/829/2/63

Schwiterman, E. W., Kiang, N. Y., Parenteau, M. N., Harman, C. E., DasSarma, S., Fisher, T. M., et al. (2018). Exoplanet biosignatures: A review of remotely detectable signs of life. *Astrobiology*, 18(6), 663–708. https://doi.org/10.1089/ast.2017.1729

Schwiterman, E. W., Robinson, T. D., Meadows, V. S., Misra, A., & Domagal-Goldman, S. (2015). Detecting and constraining N₂ abundances in planetary atmospheres using collisional pairs. *The Astrophysical Journal*, 810(1), 57. https://doi.org/10.1088/0004-637x/810/1/57

Shields, A. L., Ballard, S., & Johnson, J. A. (2016). The habitability of planets orbiting M-dwarf stars. *Physics Reports*, 663, 1–58. https://doi.org/10.1016/j.physrep.2016.10.003

Snellen, I. A. G., de Kok, R. J., le Poole, R., Brogi, M., & Birkby, J. (2013). Finding extraterrestrial life using ground-based high-dispersion spectroscopy. *The Astrophysical Journal*, 764(2), 182. https://doi.org/10.1088/0004-637x/764/2/182

Stüeken, E. E., Kipp, M. A., Koehler, M. C., & Buick, R. (2016). The evolution of Earth’s biogeochemical nitrogen cycle. *Earth-Science Reviews*, 160, 220–239. https://doi.org/10.1016/j.earscirev.2016.07.007

Tian, F., France, K., Linsky, J. L., Maus, P. D., & Veyettes, M. C. (2014). High stellar FUV/NUV ratio and oxygen contents in the atmospheres of potentially habitable planets. *Earth and Planetary Science Letters*, 385, 22–27. https://doi.org/10.1016/j.epsl.2013.10.024
References From the Supporting Information

Abbot, D. S., Cowan, N. B., & Ciesla, F. J. (2012). Indication of insensitivity of planetary weathering behavior and habitable zone to surface land fraction. *The Astrophysical Journal*, 756(2), 178.

Bézard, B., Fedorova, A., Bertaux, J.-L., Rodin, A., & Korabiev, O. (2011). The 1.10- and 1.18-μm nightside windows of Venus observed by SPICAV-IR aboard Venus Express. *Icarus*, 216(1), 173–183.

Canil, D., & Muehlenbachs, K. (1990). Oxygen diffusion in an Fe-rich basalt melt. *Geochimica et Cosmochimica Acta*, 54(11), 2947–2951.

Clough, S. A., Shepherd, M. W., Mlawer, E. J., De Lamarre, J. S., Iacano, M. J., Cady-Pereira, K., et al. (2005). Atmospheric radiative transfer modeling: A summary of the AER codes. *Journal of Quantitative Spectroscopy and Radiative Transfer*, 91(2), 233–244.

Costa, A., Caricchi, L., & Bagdassarow, N. (2009). A model for the rheology of particle-bearing suspensions and partially molten rocks. *Geochemistry, Geophysics, Geosystems*, 10(3), Q03010. https://doi.org/10.1029/2008GC002138

Eymet, V., Coustet, C., & Piaud, B. (2016). *kSpeak: An open-source code for high-resolution molecular absorption spectra production*. *Journal of Physics: Conference Series* Volume 676 Eurotherm Conference 105: Computational Thermal Radiation in Participating Media. *The Astronomical Journal*, 155(2), 170. https://doi.org/10.3847/1538-3881/aa8b08

Haelve, I., & Bachan, A. (2017). The geologic history of seawater pH, Science, 355(6329), 1069–1071.

Hier-Majumder, S., & Hirschmann, M. M. (2017). The origin of volatiles in the Earth's mantle. *Geochemistry, Geophysics, Geosystems*, 18(8), 3078–3092. https://doi.org/10.1002/2017GC006937

Hirschmann, M. M. (2000). Mantle solidus: Experimental constraints and the effects of peridotite composition. *Geochemistry, Geophysics, Geosystems*, 1, 1042.

Kadoya, S., Krissansen-Totton, J., & Catling, D. C. (2020). Probable cold and alkaline surface environment of the Hadean Earth caused by impact ejecta weathering. *Geochemistry, Geophysics, Geosystems*, 21(1), e2019GC008734. https://doi.org/10.1029/2019GC008734

Kite, E. S., Manga, M., & Gaído, E. (2009). Geodynamics and rate of volcanism on massive Earth-like planets. *The Astrophysical Journal*, 700(2), 1732.
Komacek, T. D., & Abbott, D. S. (2016). Effect of surface-mantle water exchange parameterizations on exoplanet ocean depths. *The Astrophysical Journal*, 832(1), 54.

Kress, V. C., & Carmichael, I. S. (1991). The compressibility of silicate liquids containing Fe$_3$O$_4$ and the effect of composition, temperature, oxygen fugacity and pressure on their redox states. *Contributions to Mineralogy and Petrology*, 108(1–2), 82–92.

Lécuyer, C., & Ricard, Y. (1999). Long-term fluxes and budget of ferric iron: Implication for the redox states of the Earth’s mantle and atmosphere. *Earth and Planetary Science Letters*, 165(2), 197–211.

Ma, Q., & Tipping, R. (1992). A far wing line shape theory and its application to the foreign-broadened water continuum absorption. III. *The Journal of Chemical Physics*, 97(2), 818–828.

Marrero, T. R., & Mason, E. A. (1972). Gaseous diffusion coefficients. *Journal of Physical and Chemical Reference Data*, 1(1), 3–118.

Ma, Q., & Tipping, R. (1992). A far wing line shape theory and its application to the foreign-broadened water continuum absorption. III. *The Journal of Chemical Physics*, 97(2), 818–828.

Nakayama, A., Kodama, T., Ikoma, M., & Abe, Y. (2019). Runaway climate cooling of ocean planets in the habitable zone: A consequence of seafloor weathering enhanced by melting of high-pressure ice. *Monthly Notices of the Royal Astronomical Society*, 488(2), 1580–1596.

Nimmo, F. (2007). Energetics of the core. *Treatise on Geophysics*, 8, 31–65.

Nimmo, F., & McKenzie, D. (1998). Volcanism and tectonics on Venus. *Annual Review of Earth and Planetary Sciences*, 26(1), 23–51.

O’Rourke, J. G., & Stevenson, D. J. (2016). Powering Earth’s dynamo with magnesium precipitation from the core. *Nature*, 529(7586), 387–389.

Owen, J. E. (2019). Atmospheric escape and the evolution of close-in exoplanets. *Annual Review of Earth and Planetary Sciences*, 47, 67–90.

Owen, J. E. (2019). Atmospheric escape and the evolution of close-in exoplanets. *Annual Review of Earth and Planetary Sciences*, 47, 67–90.

Papale, P. (1997). Modeling of the solubility of a one-component H$_2$O or CO$_2$ fluid in silicate liquids. *Contributions to Mineralogy and Petrology*, 126(3), 237–251.

Schwieterman, E. W., Reinhard, C. T., Olson, S. L., Harman, C. E., & Lyons, T. W. (2019). A limited habitable zone for complex life. *The Astrophysical Journal*, 878(1), 19.

Stamnes, K., Tsay, S. C., Wiscombe, W., & Jayaweera, K. (1988). Numerically stable algorithm for discrete-ordinate-method radiative transfer in multiple scattering and emitting layered media. *Applied Optics*, 27(12), 2502–2509.

Turcotte, D. L., & Schubert, G. (2002). *Geodynamics*. Cambridge University Press.

Veeder, G. J., Davies, A. G., Matson, D. L., Johnson, T. V., Williams, D. A., & Radebaugh, J. (2012). Io: Volcanic thermal sources and global heat flow. *Icarus*, 219(2), 701–722.

von Gehlen, K. (1992). Sulfur in the Earth’s mantle – A review. In K. Schidowski, S. Golubic, M. M. Kimberley, & D. M. McKirdy (Eds.), *Early organic evolution* (pp. 359–366). Springer.

Zahnle, K. J., & Kasting, J. F. (1986). Mass fractionation during transonic escape and implications for loss of water from Mars and Venus. *Icarus*, 68(3), 462–480.

Zahnle, K. J., Lupu, R., Dobrovolskis, A., & Sleep, N. H. (2015). The tethered moon. *Earth and Planetary Science Letters*, 427, 74–82.

Zeebe, R. E., & Wolfert, P. (2003). A simple model for the CaCO$_3$ saturation state of the ocean: The “Strangelove,” the “Neritan,” and the “Cretan” Ocean. *Geochemistry, Geophysics, Geosystems*, 4(12), 1104. https://doi.org/10.1029/2003GC000538