A Volcanic Hydrogen Habitable Zone

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Abstract

The classical habitable zone (HZ) is the circular region around a star in which liquid water could exist on the surface of a rocky planet. The outer edge of the traditional N2–CO2–H2O HZ extends out to nearly ∼1.7 au in our solar system, beyond which condensation and scattering by CO2 outstrips its greenhouse capacity. Here, we show that volcanic outgassing of atmospheric H2 can extend the outer edge of the HZ to ∼2.4 au in our solar system. This wider volcanic-hydrogen HZ (N2–CO2–H2O–H2) can be sustained as long as volcanic H2 output offsets its escape from the top of the atmosphere. We use a single-column radiative-convective climate model to compute the HZ limits of this volcanic hydrogen HZ for hydrogen concentrations between 1% and 50%, assuming diffusion-limited atmospheric escape. At a hydrogen concentration of 50%, the effective stellar flux required to support the outer edge decreases by ∼35%–60% for M–A stars. The corresponding orbital distances increase by ∼30%–60%. The inner edge of this HZ only moves out ∼0.1%–4% relative to the classical HZ because H2 warming is reduced in dense H2O atmospheres. The atmospheric scale heights of such volcanic H2 atmospheres near the outer edge of the HZ also increase, facilitating remote detection of atmospheric signatures.

Key words: astrobiology – planets and satellites: atmospheres – planets and satellites: terrestrial planets

1. Introduction

The habitable zone (HZ) is the circular region around a star(s) in which liquid water could be stable on a rocky planet’s surface (e.g., Kasting et al. 1993; Haghighipour & Kaltenegger 2013; Kaltenegger & Haghighipour 2013) and possible atmospheric biosignatures may be detected (e.g., Kaltenegger et al. 2010). The classical N2–CO2–H2O HZ is defined by the greenhouse effect of CO2 and H2O vapor. The outer edge, the CO2 maximum greenhouse limit, is defined as the distance beyond which condensation and scattering by CO2 outstrips its greenhouse capacity. The inner edge is defined where mean surface temperatures exceed the critical point of water (∼647 K, 220 bar), triggering a runaway greenhouse that leads to rapid water loss to space on very short timescales (see Kasting et al. 1993 for details). Additional greenhouse gases could further extend the HZ. It had been suggested that young orphan planets unbound to their host stars could accrete massive primordial hydrogen envelopes (Stevenson 1999). Models show that a super-Earth with a 40 bar primordial hydrogen atmosphere could maintain above-freezing mean surface temperatures out to 10 au from a G-type star (Pierrehumbert & Gaidos 2011). The potency of this greenhouse effect arises from collision-induced absorption (CIA) caused by self-broadening from H2–H2 collisions, allowing H2 to function noncondensibly out to great distances (Pierrehumbert & Gaidos 2011). For comparison, condensation and scattering effects at high CO2 partial pressures limit the outer edge of the classical HZ in our solar system to ∼1.7 au (e.g., Kasting et al. 1993).

However, maintaining large amounts of primordial hydrogen over geological timescales is difficult (Pierrehumbert & Gaidos 2011). Hydrogen is a very light greenhouse gas, and without a continuous source, hydrodynamic escape to space would strip a 5 Earth-mass HZ rocky planet with 50 bars of primordial atmospheric H2 within just a few million years. Primordial hydrogen lifetimes in atmospheric models only extend to ∼100 million years at H2 partial pressures of several hundred bar (Wordsworth 2012).

Another way to generate atmospheric hydrogen on terrestrial planets is through volcanism. Paleoclimate studies of early Earth and Mars show that volcanism could have outpaced escape of H2 on both planets, maintaining clement conditions with modest H2 partial pressures below 1 bar (Wordsworth & Pierrehumbert 2013; Ramirez et al. 2014). Reduced mantle conditions could have favored enhanced outgassing of H2 over longer timescales. In these models, hydrogen is not the major atmospheric constituent and is continuously replenished by volcanism that offsets losses to space. In these volcanic hydrogen atmospheres, it is foreign-broadening by the remaining background atmosphere that excites roto-translation bands within the hydrogen, increasing greenhouse warming in spectral regions where CO2 and H2O absorb poorly (Ramirez et al. 2014).

In this work, we assess the impact that foreign-broadening of volcanic hydrogen in a planet’s atmosphere has on the classical N2–CO2–H2O HZ (Kasting et al. 1993; Kopparapu et al. 2013). We derive the limits of this volcanic hydrogen HZ (N2–CO2–H2O–H2) for host stars with T_eff ranging from 2600 to 10,000 K.

2. Methodology

2.1. Climate Modeling Procedures

As in previous studies of the HZ (e.g., Kopparapu et al. 2013; Ramirez & Kaltenegger 2014, 2016), we used a 1D radiative-convective climate model to compute HZ boundaries for stars of stellar effective temperatures (T_eff) ranging from 2600 to 10,000 K (Ramirez & Kaltenegger 2016). The model uses correlated-k coefficients to parameterize absorption by H2O and CO2 across 38 solar intervals and 55 infrared intervals (e.g., see Kopparapu et al. 2013 for details). We model CO2–H2 CIA using recently computed ab initio
calculations (Wordsworth et al. 2017), CO₂—CO₂ CIA using established parameterizations (Gruszka & Borysow 1997, 1998; Baranov et al. 2004), self-broadening of H₂—H₂ pairs (Borysow 2002), and N₂ foreign-broadening (see Ramirez et al. 2014 for details).

We use inverse calculations, where the surface temperature is specified and the solar flux required to maintain it is calculated, to determine the limits of the HZ (following, e.g., Kasting et al. 1993; Kopparapu et al. 2013). For the outer edge HZ calculations, the mean surface temperature is fixed at 273 K, above which liquid water is stable on a planetary surface. Stratospheric temperatures are fixed at 154 K, which is representative for a planet near the outer edge of the HZ (e.g., Kopparapu et al. 2013). For the inner edge, the stratospheric temperature is 200 K and the surface temperature is gradually increased from a starting temperature of 200 K, which simulates pushing the planet closer and closer to the star until a runaway greenhouse is triggered (e.g., Kasting et al. 1993).

The input of hydrogen from volcanic sources in our models is balanced by its escape to space. Actual hydrogen escape rates will be planet-specific; therefore, we assume that hydrogen escapes at the diffusion limit, which is the fastest that H can escape at the concentrations considered (e.g., Walker 1977). Thus, our results may overestimate hydrogen escape and thus provide a lower bound on H concentrations. As with previous HZ studies, we assume planets support plate tectonics, including a carbonate-silicate cycle capable of maintaining habitability over long timescales (e.g., Kasting et al. 1993; Ramírez & Kaltenegger 2016). We also assume highly reduced mantle conditions on terrestrial planets, as suggested from meteoritic evidence from early Mars (e.g., Grott et al. 2011). Hydrogen sources and sinks need to balance to support a continuous hydrogen concentration. Our model planet has the same outgassing rates per unit area as those on our present-day planet based on the inference that the heat flux of early Mars could have been similar to that of modern Earth (Montes & Zuber 2003; Ramírez et al. 2014). This allows us to constrain typical H₂ concentrations for reducing mantles of outer edge planets to ~1%–30% (Ramírez et al. 2014; Batalha et al. 2015). For comparison, we include one extreme concentration of 50% (assuming even higher outgassing rates or less efficient hydrogen escape) intended to simulate potentially higher outgassing rates (e.g., on hypothetically planets with unusually high geothermal fluxes).

Each of our model atmospheres contain 1 bar of N₂ and H₂ concentrations of 1%, 5%, 10%, 20%, 30%, and 50%, respectively, of the dry (N₂+CO₂+H₂) atmosphere. Oxygen has only a weak greenhouse effect and is replaced by N₂ in these calculations (following Kasting et al. 1993). For the outer edge calculations we vary the CO₂ partial pressure from 1 × 10⁻² to 34.7 bar, the saturation CO₂ partial pressure at 273 K. CH₄ is not expected to build up to appreciable concentrations in CO₂–H₂ atmospheres because of its short photochemical lifetime when H₂ is not the major constituent (Bains et al. 2014). Recent work shows that CH₄ would likely oxidize to CO₂ in warm CO₂–H₂ atmospheres (Batalha et al. 2015).

At the inner edge, our volcanic H₂ atmospheres contain 1 bar of N₂ and 330 ppm CO₂ (following Kasting et al. 1993). As in the case of the outer edge, we model H₂ concentrations that are 1%, 5%, 10%, 20%, 30%, and 50% of the dry (N₂+CO₂+H₂) atmosphere. We gradually increase planetary surface temperatures from 200 K to the critical point of water (647.1 K, 220.6 bars H₂O), which leads to the runaway greenhouse and rapid evaporation of the surface water inventory (e.g., Kasting et al. 1993). The high atmospheric H₂O mole fraction at the inner edge reduces the warming effect of H₂. Note that the H₂ amount is scaled to the dry atmosphere because volcanism is not expected to increase as H₂O vapor mixing ratios increase.

In keeping with previous work (see, e.g., Kopparapu et al. 2013), all model atmospheres are fully saturated: a moist CO₂ adiabat is followed in the upper atmosphere for the outer edge calculations. A moist H₂O adiabat is followed in the upper atmosphere for the inner edge model calculations. We also employ six solar zenith angles and a surface albedo of 0.31 to mimic the combined clouds and surface reflection, which reproduces the mean surface temperature for present Earth (ibid). Because there is no self-consistent cloud model for 1D HZ studies (see Zsom et al. 2012), surface and cloud albedo are held constant for all simulations (following previous 1D HZ studies; see, e.g., Kasting et al. 1993; Kopparapu et al. 2013), even though cloud feedback is expected to change for atmospheric compositions different from those of present-day Earth. However, our understanding of clouds is still evolving and their applicability to water-rich atmospheres (i.e., inner edge of the HZ) is not fully understood yet. Note that overall planetary albedo changes as the atmospheric composition is varied.

3. Results

Adding volcanic hydrogen moves both the inner and the outer edges of the classical HZ outward and widens it. The incident stellar fluxes (S_\text{eff}) corresponding to the outer edges of the classical and new volcanic H₂ HZ are shown in Figure 1. For comparison, an alternative HZ limit that is not based on atmospheric models (i.e., classical HZ), but on empirical observations of our solar system, is also shown in Figures 1–2. The inner edge of this empirical HZ is defined by the stellar flux received by Venus when we can exclude the possibility that it had standing water on the surface (about 1 Gyr ago), equivalent to a stellar flux of S_\text{eff} = 1.77 (Kasting et al. 1993). The outer edge is defined by the stellar flux that Mars received.

![Figure 1. Effective stellar temperature vs. incident stellar flux (S_\text{eff}) for the outer edge. The CO₂ maximum greenhouse limit (dashed) is shown along with the empirical outer edge (solid black) and outer edge limits containing 5%, 10%, 20%, 30%, and 50% H₂ (red solid).](image-url)
The incident flux in both figures is normalized to that received by Earth’s orbit (∼1360 W m⁻²), S₀.

The incident stellar flux that is needed to maintain liquid water surface temperatures at the outer edge of our solar system decreases from 0.3576 to 0.267 (25% decrease) and to 0.201 (44% decrease) with 5% and 30% H₂, respectively (Figure 1). This corresponds to increases in the orbital distance of the maximum greenhouse limit of CO₂ at the outer edge HZ from 1.67 au to 1.94 au and 2.23 au, respectively (16% and 34% increases, respectively; Figure 2). With 50% H₂ the limits extend farther, corresponding to an Sₑff of 0.172 (52% decrease) and an outer edge distance of 2.4 au (44% increase) in our own solar system. Overall, the Sₑff required to support the outer edge decreases by a maximum of ∼35%–60% (at 50% H₂) for M–A star types, resulting in respective distance increases of ∼30% to 60% (Figure 2).

The volcanic hydrogen HZ boundaries can be calculated using Equation (1) with constants from Tables 1 and 2:

\[ S_{\text{eff}} = S_{\text{eff(Sun)}} + aT_*^2 + bT_*^3 + cT_*^4 + dT_*^5, \]  

(1)

where \( T_* \) is equal to \( T_{\text{eff}} = 5780 \text{ K} \) and \( S_{\text{eff(Sun)}} \) are the fluxes computed for our solar system (see Tables 1–2). The corresponding orbital distances of the HZ can be calculated using Equation (2) (Figure 2):

\[ d = \frac{L/L_{\text{Sun}}}{S_{\text{eff}}}, \]  

(2)

where \( L/L_{\text{Sun}} \) is the stellar luminosity in solar units and \( d \) is the orbital distance in astronomical units. Planetary mass that is linked to planetary model surface pressure via gravity and planetary radius have only a small effect on the HZ limits (Kopparapu et al. 2014). Although the inner edge of the HZ is also affected by the addition of H₂, the runaway greenhouse limit distance (e.g., Kasting et al. 1993) only moves outward by ∼0.1% to 4% for 1% H₂ and 50% H₂ (Figure 2), respectively, because of the dense water vapor atmosphere at the inner edge. At temperatures above ∼300 K, water vapor absorption begins to mask hydrogen CIA, decreasing its effectiveness.

The greenhouse effect of CO₂ increases with higher pCO₂ to a maximum value (Figure 3), corresponding to the maximum

![Figure 2. Effective stellar temperature versus (left) incident stellar flux (Sₑff) and (right) orbital distance for the classical (dashed), empirical (solid black), and volcanic hydrogen (solid red) habitable zone. The red curves contain 30% and 50% H₂, respectively. As shown in Kopparapu et al. (2013), Earth appears near the classical inner edge because of the generic model assumption of 100% relative humidity. With subsaturated air, Earth is well within the classical HZ (e.g., Leconte et al. 2013).](image-url)
greenhouse effect. However, as the hydrogen concentration increases, the CO₂ partial pressures associated with the outer edge occur at lower $S_{\text{eff}}$ and larger orbital distances. This occurs because cooling from both CO₂ condensation and scattering eventually outpace the combined greenhouse effect of CO₂ and H₂ farther from the host star.

About 8 bars of CO₂ are required to sustain surface temperatures above freezing at the CO₂ maximum greenhouse limit of the classical HZ in our solar system (Kasting et al. 1993; Kopparapu et al. 2013). When hydrogen is added to the planetary atmosphere, the CO₂ partial pressure required at the outer edge of the volcanic H₂ HZ decreases to 6 bar with 1% H₂, and to 1 bar at 30% H₂ (Figure 3). The corresponding H₂ partial pressures (with 1 bar N₂) are 0.071 bar and 0.857 bar, respectively. At 50% H₂, the CO₂ and H₂ partial pressures at the outer edge of the volcanic hydrogen HZ are 0.5 and 1.5 bars, respectively.

The stellar energy distribution of host stars also determines what CO₂ concentration the maximum CO₂ greenhouse effect is reached. For cooler stars, increased near-infrared absorption and reduced scattering lower this CO₂ concentration (e.g., Kasting et al. 1993). For a cool M6 star ($T_{\text{eff}} \sim 3000 \text{ K}$), the CO₂ partial pressure at the outer edge of the classical HZ is 15 bar. Adding H₂ reduces the required amount of CO₂ at the outer edge to 9 bar and 1 bar for 1% and 30% H₂, respectively. The corresponding H₂ partial pressures are 0.1 bar and 0.857 bar.

Atmospheric scale-height increases with the addition of hydrogen. For example, adding 30% H₂ to the atmosphere of a habitable planet located in our solar system’s outer edge (1 bar N₂ and 8 bar CO₂) increases its atmospheric scale-height by over a factor of 1.5 due to the decreased atmospheric mean molecular mass, assuming a similar temperature structure. In addition, such H₂-rich planets at the outer edge of the HZ also have less CO₂ in their atmospheres, reducing the optical density as compared to those CO₂ atmospheres that do not contain hydrogen. These two effects could facilitate the detection of atmospheric spectral features with the next generation of telescopes, making such planets very interesting targets for JWST and the E-ELTs in the search for life.

**Figure 3.** Dependence of the incident stellar flux ($S_{\text{star}}$) on atmospheric CO₂ partial pressure (pCO₂) for the solar system outer edge. The pCO₂ values corresponding to the maximum greenhouse for atmospheres with hydrogen (red curves) and without (dashed curve) are shown as vertical lines for 0% to 50% H₂ (see the text).

4. Discussion

Although the present-day mantle of Earth is relatively oxidized, it is generally assumed that the mantles in terrestrial planets, including Earth, start reduced (i.e., oxygen-poor), which may lead to conditions that promote high levels of H₂ outgassing. One of the main arguments for a reduced early mantle is that Earth (and possibly other terrestrial planets) must have entered a magma ocean stage during accretion, which would have acted as a surficial oxygen sink (e.g., Scaillet & Gaillard 2011). Afterward, planetary mantles are thought to become oxidized over time, either gradually or in a stepwise fashion (Wood et al. 2006). Mantle oxidation on Earth may have occurred rather quickly, perhaps only ~100 million years after core–mantle differentiation (Trail et al. 2011).

Mantle oxidation may have taken longer on Mars though. High H₂ outgassing on early Mars is inferred from Martian meteorites suggesting a highly reduced ancient Martian mantle (Grott et al. 2011; Stanley et al. 2011). If early Mars was kept warm by atmospheric H₂ up until ~3.6 Ga, this would suggest reducing mantle conditions had persisted for ~1 billion years. Supporting this view, geochemical results from the ~4 billion year old meteorites, ALH84001 and NWA Black Beauty, infer a reduced mantle for at least ~0.5 billion years (e.g., Grott et al. 2011).

These solar system examples warrant some discussion of how mantle oxidation may generally proceed on planets. One idea suggests lower mantle pressures of larger terrestrial bodies (e.g., Earth) are sufficiently high to convert iron(II) oxide to iron metal and iron(III) oxide, oxidizing the mantle during accretion (Wade & Wood 2005). According to this notion, smaller planets like Mars would not have generated internal pressures sufficiently high for mantle oxidation, in agreement with the meteorite evidence. This may suggest that atmospheric H₂ may not easily build up on more massive planets. However, this may happen in a number of ways. First, these calculations assume that H₂ escapes at the diffusion limit, but H₂ concentrations can potentially build up if escape rates (including hydrodynamic and Jean’s) are lower than this limit, as had been suggested for the early Earth (Tian et al. 2005). Hydrogen retention will also be favored on planets with stronger magnetic fields or on those with higher outgassing rates per unit area.

Moreover, it is possible that H₂-rich atmospheres may be longer-lived on super-Earths. First, their enhanced gravity results in reduced H₂ escape rates, promoting hydrogen retention (e.g., Pierrehumbert & Gaidos 2011). Second, mantle oxidation may be slowed as a result, increasing the lifetime of reduced mantle conditions. Larger planets may have higher outgassing rates, possibly leading to a feedback in which higher outgassing rates enhance weathering rates and vice versa (e.g., Heller 2015). Plus, interior heat is retained longer on super-Earths, ensuring that plate tectonics and volcanism both last (ibid).

Hydrogen retention is also favored for outer edge planets. Hamano et al. (2013) suggest that selective hydrodynamic escape of H over O during accretion could lead to early mantle oxidation. However, outer edge planets would be less susceptible to this effect because corresponding EUV fluxes are much lower than in the inner solar system. Given that hydrodynamic escape rates are $\propto 1/(d(\text{au})^2)$, and assuming energy-limited escape (e.g., Ramirez & Kaltenegger 2014), H escape rates (all else equal) on a young accreting Earth-mass
planet orbiting the early Sun at 1.7 au would have been \( \sim \) a factor of \( \sim 3-6 \) lower than on early Earth or Venus. These loss rates would decrease further by 1/30–1/60 on a 10 Earth-mass super-Earth. It is also possible that accreting planets incorporate more hydrogen because the protoplanetary disk itself is enriched in the element. In addition, because stellar fluxes are lower for these distant planets, magma ocean durations should decrease as well, further weakening escape (Hamano et al. 2013). All of these factors could more than offset the factor of \( \sim 15-40 \) relatively higher \( \text{H}_2 \) outgassing rates inferred for a highly reduced mantle (Ramirez et al. 2014). Thus, sizable hydrogen inventories on larger terrestrial planets located near the outer edge may be sustainable as long as mantle conditions support conditions that favor hydrogen retention over escape to space.

Certain atmospheric spectral features, including \( \text{N}_2 \)O and \( \text{NH}_3 \), which can, but do not have to be, produced biotically could be detected in \( \text{H}_2 \)-dominated atmospheres (Seager et al. 2013; Bains et al. 2014). Such volcanic hydrogen atmospheres may also be able to evolve methane-based photosynthesis (Bains et al. 2014). Distinguishing biosignatures from abiotic sources in such atmospheres will be challenging. \( \text{NH}_3 \) can be formed abiotically through reaction of \( \text{N}_2 \) and \( \text{H}_2 \) in hydrothermal vents on planets with reducing mantles (Kasting et al. 2014). \( \text{N}_2 \)O can be formed in a number of ways including through atmospheric shock from meteoritic fall-in, lightning, UV radiation (e.g., Ramirez 2016), and through solar flare interactions with the magnetosphere (Airapetian et al. 2016). Thus, future biosignature studies should focus on modeling biotic and abiotic sources for these gases in thin volcanic hydrogen atmospheres.

Lastly, earlier studies suggesting that backscattering from \( \text{CO}_2 \) clouds could move the outer edge beyond 2 au (Mischna et al. 2000) is challenged by new work indicating that the greenhouse effect of \( \text{CO}_2 \) clouds may be greatly overstated (Kitzmann 2016, 2017). Even should warming from \( \text{CO}_2 \) clouds be relatively ineffective, volcanically outgassed \( \text{H}_2 \) in some stellar systems may compensate and keep the \( \text{HZ} \) similarly wide for geologically significant timescales.

5. Conclusion

We have calculated the limits of the volcanic \( \text{H}_2 \) \( \text{HZ} \) for \( \text{H}_2 \) concentrations ranging from 1% to 50%. We find that the addition of 30% \( \text{H}_2 \) can extend the outer edge in our solar system to \( \sim 2.23 \) au (2.4 au with 50% \( \text{H}_2 \)). The orbital distance of the outer edge of the volcanic hydrogen \( \text{HZ} \) moves outward by as much as \( \sim 60\% \) as compared to the classical \( \text{HZ} \) for A to M-type main-sequence stars. The inner edge only moves out by \( \sim 0.1\% \) to 4%.

Reduced mantle conditions that promote hydrogen out-gassing on small terrestrial planets may be maintained for long timescales (>0.5 Gyr). Atmospheric \( \text{H}_2 \) may be retained on larger terrestrial planets, including super-Earths, through higher gravity and potentially stronger magnetic fields. \( \text{H} \) retention is also generally favored in planets with higher outgassing rates, and if \( \text{H}_2 \) escapes more slowly than the diffusion limit, Hydrodynamic escape rates are also lower for planets farther away from the host star, mitigating the selective escape of \( \text{H} \) over O. Generally, hydrogen-rich atmospheres composed of \( \text{N}_2 \), \( \text{CO}_2 \), and \( \text{H}_2 \text{O} \) can be maintained as long as volcanic outgassing of \( \text{H}_2 \) outpaces its escape.

Moreover, the addition of \( \text{H}_2 \) increases the atmospheric scale-height making super-Earths near the volcanic \( \text{H}_2 \) \( \text{HZ} \) outer edge interesting observational targets. Near-future missions like \( \text{JWST} \) and the ELTs could potentially detect atmospheric features, including biosignatures, on such planets.

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