A Volatile-poor Formation of LHS 3844b Based on Its Lack of Significant Atmosphere

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Abstract

Exoplanet discoveries have reached into the realm of terrestrial planets that are becoming the subject of atmospheric studies. One such discovery is LHS 3844b, a 1.3 Earth radius planet in a 0.46 day orbit around an M4.5-5 dwarf star. Follow-up observations indicate that the planet is largely devoid of substantial atmosphere. This lack of significant atmosphere places astrophysical and geophysical constraints on LHS 3844b, primarily the degree of volatile outgassing and the rate of atmosphere erosion. We estimate the age of the host star as 7.8 ± 1.6 Gyr and find evidence of an active past comparable to that of Proxima Centauri. We use geodynamical models of volcanic outgassing and atmospheric erosion to show that the apparent lack of atmosphere is consistent with a volatile-poor mantle for LHS 3844b. We show the core is unlikely to host enough C to produce a sufficiently volatile-poor mantle, unless the bulk planet is volatile-poor relative to Earth. While we cannot rule out a giant impact stripping LHS 3844b’s atmosphere, we show that this mechanism would require significant mantle stripping, potentially leaving LHS 3844b as an Fe-rich “super-Mercury.” Atmospheric erosion by smaller impacts is possible, but only if the planet has already begun degassing and is bombarded by 10^3 impactors of radius 500–1000 km traveling at escape velocity. We discuss formation and migration scenarios that could account for a volatile-poor origin, including the potential for an unobserved massive companion planet. A relatively volatile-poor composition of LHS 3844b suggests that the planet formed interior to the system snow line.

Unified Astronomy Thesaurus concepts: Exoplanets (498); Stellar activity (1580); Exoplanet atmospheres (487); Planetary interior (1248); Planet formation (1241)

1. Introduction

Exoplanetary science has advanced to a regime where terrestrial planets are routinely discovered orbiting other stars. For example, the Kepler mission discovered several hundred terrestrial planets during the primary four years of observations (Borucki 2016), of which tens were found to lie within their star’s habitable zone (Kane et al. 2016). Similar terrestrial planet discoveries are now continuing through the use of the Transiting Exoplanet Survey Satellite (TESS), launched into Earth orbit in 2018 (Ricker et al. 2015). It is expected that the relative brightness of TESS host stars will allow the discovered planets to become prime targets for atmospheric characterization using follow-up facilities, such as the James Webb Space Telescope (Kempton et al. 2018; Ostberg & Kane 2019). Such observations will allow tests of composition and atmospheric erosion scenarios (Owen 2019; Rodríguez-Mozos & Moya 2019). Such erosion scenarios are considered particularly relevant to low-mass stars that can have extended periods of high activity, such as the case of TRAPPIST-1 (Roettenbacher & Kane 2017; Dong et al. 2018).

One of the very early discoveries announced using data from the TESS mission was the detection of a planet orbiting the star LHS 3844 (Vanderspeck et al. 2019) with an orbital period of ~11 hr. The planet is likely terrestrial, with a radius of \( R_p = 1.303 \pm 0.022 \, R_{\text{Earth}} \), although no strong constraints on the planet mass have yet been established. The importance of the planet was raised significantly by Kreidberg et al. (2019b), who reported observations of the LHS 3844 system using the Spitzer space telescope. Their analysis of the Spitzer data indicated that the planet does not have a thick atmosphere, down to a limit of 10 bar, consistent with severe atmospheric erosion of the primary atmosphere. It remains to be seen whether similar atmospheric erosion effects are common among terrestrial planets around low-mass stars, and how these effects depend on planetary mass and age.

The build up of a secondary atmosphere is a coupled astronomical and geological process. Geologically, melting in the shallow subsurface of the planet releases volatile gases to the atmosphere (e.g., Holland 1984; Zhang 2014; Foley & Driscoll 2016; Noack et al. 2017; Tosi et al. 2017), while atmospheric erosion due to stellar activity removes this atmosphere over time (Sakuraba et al. 2019). The special case of LHS 3844b currently having little-to-no atmosphere then allows us to simultaneously constrain both the atmospheric environment the planet has experienced as well as aspects of its geological evolution and composition.

Here, we present the results of a combined stellar and geophysical study that aims to provide context for the observed lack of thick atmosphere for LHS 3844b. In Section 2, we provide estimates for the age and activity of the host star based on known stellar parameters. In Section 3, we calculate a geologically motivated planetary mass and density, and place the incident flux received by the planet in the context of other similar exoplanets. Section 4 provides a detailed description of our interior model and method for estimating potential degassing timescales and atmospheric erosion scenarios. The results of our model calculations, including the time evolution
of atmospheric pressure, are provided in Section 5. In Section 6, we discuss the implications of the geodynamical models, core and mantle volatile inventory, and stellar erosion factors for the formation and migration scenarios of the planet relative to the snow line. We also include a detailed description of the effect of potential small and giant impactors on the atmospheric evolution and evaluate the likelihood of impacts as the source of the observed constraints for the LHS 3844b atmosphere. We provide concluding remarks in Section 7, including future implications and applications of this work.

2. Host Star Properties

We adopt the accumulated stellar parameters provided by Vanderspek et al. (2019), which include mass, radius, and effective temperature of $M_* = 0.151 \pm 0.014 M_{\odot}$, $R_* = 0.189 \pm 0.006 R_{\odot}$, and $T_{\text{eff}} = 3036 \pm 77$ K, respectively. Vanderspek et al. (2019) further provide a spectral type of M4.5 or M5, and estimate a stellar rotation period of $128 \pm 24$ days based on time series photometry. For this study, we further estimate the age and the activity properties of the star.

2.1. Age

Gyrochronology is a useful tool for constraining stellar ages by combining observed rotation periods with other intrinsic stellar properties (Barnes 2007; Angus et al. 2019a). However, the methodology becomes increasingly uncertain for very low-mass stars where rotation rates are poorly calibrated with age estimates (Angus et al. 2019b; Gallet & Delorme 2019). In order to use gyrochronology, stars are required to spin down over time due to angular momentum loss. This angular momentum loss has been shown to be strongly tied to the stellar open flux, which is the magnetic flux found in the open field lines of the stellar magnetic field that carries winds away from the star (Mestel & Spruit 1987). For stars with convective outer envelopes, this open flux can be estimated to be a dipolar field (e.g., Petit et al. 2008). However, low-mass M dwarfs that are fully convective can have their open flux estimated as either a dipolar field or a more complex field structure (e.g., Donati et al. 2008). Even knowing the structure of the magnetic field does not allow observers to understand the nature of the spin-down of these low-mass stars (See et al. 2017).

Newton et al. (2016) used photometry from the MEarth Project to obtain rotation periods and combined that information with a proper motion survey (Lépine & Shara 2005) to estimate the ages of low-mass stars based on how their rotation period correlates to where in the galaxy the star is found. LHS 3844 has a reported rotation period of $128 \pm 24$ days (Vanderspek et al. 2019) from ground-based MEarth monitoring, but was not included in the Newton et al. (2016) sample. For stars with rotation periods greater than 70 days, Newton et al. (2016) estimated that stars have an average age of $5.1^{+0.3}_{-0.2}$ Gyr. Their sample included 28 stars with an average rotation period of 106.2 days. Engle & Guinan (2018) provided a rotation-age relationship for M dwarfs, making use of ten years of ground-based photometry and the stars’ membership or probable membership in a cluster or group. The relationship for M dwarfs with spectral types M2.5-M6, within which LHS 3844 falls, suggests an age of $7.8 \pm 1.6$ Gyr, which is consistent with the Newton et al. (2016) estimate and provides the best age estimate presently available for a late-M field star like LHS 3844.

2.2. Activity

We calculated the Rossby number, or the ratio of stellar rotation period to the convective turnover time, $R_o \equiv P_{\text{rot}}/T_{\text{conv}}$, for LHS 3844 using Equations (10) and (11) from Wright et al. (2011). Equation (10) depends only upon $V - K_s$ color and Equation (11) depends only upon stellar mass (in terms of solar masses). For these quantities, we used parameters reported by Vanderspek et al. (2019). We found that the Rossby number of LHS 3844 is approximately $R_o = 0.90 \pm 0.18$ when using Equation (10) and the appropriate photometric magnitudes, and $R_o = 1.22 \pm 0.25$ when using Equation (11) and the mass of the star (see Figure 1). Both of these values suggest that the star presently is not strongly active. A Rossby number of $R_o \leq 0.1$ indicates that a star is fully saturated in activity (e.g., Wright et al. 2011; Wright & Drake 2016; Newton et al. 2017; Wright et al. 2018). That is, faster rotation cannot increase the amount of activity observed from the star. However, as Rossby numbers increase above $R_o = 0.1$, due to increasing rotation periods, the amount of activity observed on the star will decrease. Largely due to its slow rotation period of $P_{\text{rot}} = 128 \pm 24$ days, LHS 3844 has a Rossby number significantly higher than 0.1, and the range of values calculated suggest that LHS 3844 is currently (weakly) active but not saturated. The light curve of LHS 3844 shows evidence of rotational modulation, likely due to starspots rotating in and out of view, but no evidence of other activity events, such as stellar flares.

Because the nature of angular momentum loss for fully convective stars is not well-determined, we are only able to suggest that, if LHS 3844 has spun down over time, then it was previously more active (e.g., Reiners & Basri 2008). While there is no evidence of flares in the TESS light curve, if we assume that fully convective stars spin down with the relationship given by Engle & Guinan (2018), we can use the...
activity of Proxima Centauri (hereafter Proxima Cen) to understand how LHS 3844 may have behaved in the past. Proxima Cen is an M5.5 dwarf with an age of approximately 6 Gyr and a rotation period $P_{rot} \approx 83$ days. Vida et al. (2019) found that Proxima Cen had 1.49 flares/day in two sectors of TESS data (Sectors 11 and 12) with flares ranging from $10^{30}$ to $10^{32}$ erg. They predict that larger flares on the order of $10^{34}$ erg once every two years for a star this active. While it is difficult to project stellar activity forward or backward with fully convective M dwarfs, we can use Proxima Cen as a proxy for how LHS 3844 could have behaved in the past. We therefore adopt the range of atmospheric loss rates predicted by Kreidberg et al. (2019b), who scaled the loss rate from Proxima Cen b to the value of 30–300 kg s$^{-1}$ (Dong et al. 2017).

3. Planetary Properties

3.1. Mass, Radius, and Density

In order to estimate volatile degassing rates for the planet, it is necessary to determine fundamental planetary properties, including the mean (bulk) density. The radius of the planet, provided by Vanderspek et al. (2019), is $1.303 \pm 0.022$ $R_{\oplus}$. We utilize the mass–radius–composition methodology of Unterborn & Panero (2019) in order to estimate the mass of LHS 3844b. Knowing the planet lacks any significant volatile atmosphere, we assume the planet is made entirely of an FeO-free silicate mantle and pure liquid-Fe core. Mantle phase equilibria and core radius fractions are calculated using the mass–radius–composition solver, ExoPlex (Unterborn & Panero 2019). We adopt the thermodynamic equation-of-state data of Stixrude & Lithgow-Bertelloni (2011) to calculate mantle phase equilibria and use the fourth-order Birch–Murnaghan equation of state for liquid Fe of Anderson & Ahrens (1994). These models accurately reproduce the Earth (Unterborn et al. 2016), provide more robust estimates for masses of individual planets than do empirical models (e.g., Zeng et al. 2016), and are also wholly self-consistent with mineral physics experimental data (Unterborn & Panero 2019).

There are no estimates of the bulk composition of LHS 3844b, whether directly through measurement of its density or indirectly from estimates of host-star composition. As the mass of a rocky planet (for a given radius) is most sensitive to the relative size of the central Fe core, as defined by its core mass fraction (CMF), we choose two end-member compositions of the modeled planet’s bulk Fe/Mg of 0.6 and 1.5. This equates to CMFs of 0.25 and 0.45, respectively (n.b., the Earth’s CMF is 0.33). These end members encompass 80% of all measured stellar Fe/Mg within the Hypatia catalog (Hinkel et al. 2014; Unterborn & Panero 2019).

From this modeling, we estimate the mass of LHS 3844b to be $2.4 M_{\oplus}$ for the small core case, $2.5 M_{\oplus}$ for the Earth-like core case, and $2.9 M_{\oplus}$ for the large core case. These equate to core radius fractions of 0.46, 0.55, and 0.59, respectively. We calculate the bulk planet density then varying between 6 and 7.2 g cm$^{-3}$ for the small and large core cases, respectively, making them slightly more dense than the Earth (5.5 g cm$^{-3}$). We note that the probabilistic model forecasting tool provided by Chen & Kipping (2017) estimates a mass for the planet of $2.20^{+1.07}_{-0.65} M_{\oplus}$, leading to a bulk density of $5.5^{+1.7}_{-1.0} g$ cm$^{-3}$. While this result is consistent with our model, we attribute the large uncertainties in the Chen & Kipping (2017) model result to LHS 3844b lying slightly above the boundary between their defined “Terran” and “mini-Neptune” mass boundaries. Compared to the Terran planets, there is a larger uncertainty in mass for a given radius within the mini-Neptune regime, explaining the nearly factor-of-two uncertainty in the Chen & Kipping (2017) modeled bulk density.

3.2. Incident Flux

The incident (insolation) flux environment of the planet is a key factor in the atmospheric evolution of the planet. To place LHS 3844b in context, we extracted data for the confirmed exoplanets from the NASA Exoplanet Archive (Akesson et al. 2013). The data are current as of 2020 February 16. We selected those systems whose host star mass is less than 0.5 $M_{\odot}$ and planets that either have a mass less than 10 $M_{\oplus}$ and/or a radius less than 2 $R_{\oplus}$. We used the stellar information provided, specifically the stellar luminosities and semimajor axes, to calculate the insolation flux for each planet. These data are shown in Figure 2 for all of the planets that met the above criteria. We do not include insolation uncertainties in the plot because, although stellar mass uncertainties are readily available, uncertainties on insolation and luminosities are relatively scarce, resulting in a significant diversity of insolation uncertainties (3%–30%) that would appear potentially misleading. LHS 3844b receives ~70 times the flux received by Earth and is indicated by a star near the bottom left corner of the plot. TRAPPIST-1 planets are indicated by triangles.

![Figure 2](image-url)
to disfavor plate tectonics (e.g., Lenardic & Kaula 1994; Regenauer-Lieb et al. 2001; Lenardic et al. 2008; Landuyt & Bercovici 2009; Korenaga 2010; Foley et al. 2012). As a result of stagnant lid tectonics and the lack of surface water, volcanism is expected to produce a basaltic crust covering the planet’s surface; this crustal composition is consistent with that inferred by Kreidberg et al. (2019b).

The primary volatile species outgassed by volcanism on Earth are H2O and CO2. As LHS 3844b receives too much radiation from the star for liquid water to be stable at the surface (Tian & Ida 2015), we focus on CO2. Whether oxidized species, like CO2, or more reduced species, like CO or CH4, are outgassed depends on the oxidation state of the mantle (e.g., Kasting et al. 1993; Gaillard & Scaillet 2009). Oxidation of Earth’s mantle is thought to occur just after planet formation by disproportionation in the lower mantle (e.g., Wade & Wood 2005; Wood et al. 2006). This process is expected to occur on rocky planets of Earth size or larger, as long as the mantle mineral makeup is dominated by Mg-silicates. Assuming an oxidized mantle for LHS 3844b is therefore reasonable. However, even if the mantle of LHS 3844b is more reduced than Earth’s, the same process of mantle thermal evolution and degassing that we model still determines how atmospheric mass evolves over time.

We modify the model of Foley & Smye (2018) to apply to LHS 3844b. As the exact mass and interior structure of LHS 3844b is unknown, we consider three end members as outlined in Section 3.1: a small core size (46% by radius), a large core size (59% by radius), and an Earth-like core size (55% by radius). For all cases, the planet radius is held fixed at 1.3 \( R_{\oplus} \), as radius is tightly constrained, yielding masses between 2.4 and 2.9 \( M_{\oplus} \) for the small and large core sizes, respectively. For the small core models, ExoPlex calculated the average mantle density \( \rho = 5400 \text{ kg m}^{-3} \) and gravity \( g = 14.3 \text{ m s}^{-2} \); for the large core models, \( \rho = 5360 \text{ kg m}^{-3} \) and \( g = 18.9 \text{ m s}^{-2} \); and for the Earth-like core, \( \rho = 5380 \text{ kg m}^{-3} \) and \( g = 15.7 \text{ m s}^{-2} \). As the planet cannot sustain liquid water oceans or weathering, all CO2 outgassed from the mantle is assumed to accumulate in the atmosphere; this means that there is no degassing flux from decarbonation of the crust, as there is no weathering to deposit significant stores of carbon into the crust. We also conservatively assume that melt becomes denser than the solid mantle above a pressure of 6 GPa, and thus only melt formed at pressures lower than this contributes to mantle outgassing. This chosen melt density crossover pressure is on the low end of experimental estimates (e.g., Suzuki & Ohtani 2003; Reese et al. 2007), meaning that uncertainty in the melt density will only serve to increase outgassing rates and hasten the rapid mantle outgassing our models predict (see Section 5).

We assume the mantle has a finite CO2 budget, which is set by an initial concentration of CO2 in the mantle, \( C_{\text{conc}} \). We assume a range of \( C_{\text{conc}} = 10^{-4} - 10^{-2} \text{ wt\%} \), varying from the Earth’s assumed CO2 concentration of \( \sim 10^{-4} \text{ wt\%} \) (based on an estimate of 1022 mol of CO2 in the mantle and surface reservoirs of Earth from Sleep & Zahnle 2001) to two orders of magnitude lower. This range of CO2 concentrations results in a range of total mantle CO2 budgets, \( C_{\text{tot}} \), of \( C_{\text{tot}} \approx 2.5 \times 10^{20} - 2.5 \times 10^{22} \) mol in our models. Note that a given mantle CO2 concentration results in slightly different total CO2 budgets for the three end member planet interior structures we model, as changing the core size changes the total mass of the mantle. As the results show (Section 5), our chosen range of CO2 budgets brackets the key

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**Figure 3.** Schematic of our planetary evolution model of LHS-3844b undergoing stagnant lid tectonics adapted from Foley & Smye (2018). Host star provides an age of the system over which to model the geodynamic evolution (Section 2.1), information on the atmospheric stripping rate through estimates of the XUV flux given its stellar type (Section 2.2), and the maximum core mass fraction from the measurements of the host star’s Fe/Mg ratio, assuming all the Fe is in the core, which sets the size of the convecting mantle (Section 3.1). Pink arrows represent the flux of volatile species into and out of the planet’s atmosphere. We assume there was never water on LHS-3844b, as the presence of water is necessary for sequestering CO2 into the mantle via atmospheric escape. This lack of water also prevents degassing from the crust via metamorphic reactions. Thus, the two primary fluxes for CO2 in our model for LHS-3844b’s evolution are the atmospheric escape rate (30–300 kg s\(^{-1}\)) and the rate of mantle degassing.

**4. Atmospheric Evolution**

**4.1. Atmosphere and Degassing**

After a planet has lost its primary H2/He atmosphere, the creation of a secondary atmosphere is primarily due to degassing of volatiles from the interior by mantle volcanism. The rate of volatile outgassing from the interior via mantle volcanism—and hence the total size of atmosphere that can accumulate over time—is controlled by the thermal evolution of the mantle (e.g., Tajika & Matsui 1992; Hauck & Phillips 2002; Grott et al. 2011; Tosi et al. 2017; Dom et al. 2018; Foley & Smye 2018). We use simple thermal evolution models to constrain outgassing history and the resulting atmospheric mass, based on the works of Foley & Smye (2018) and Foley (2019). Figure 3 shows a schematic of our model including those values obtained by analysis of LHS 3844 and the fluxes of CO2 between the mantle, crust, and atmosphere of LHS 3844b.

We seek to determine the combinations of planetary properties that can lead to an atmosphere size <10 bar, as observed for LHS 3844b, by its present-day age. We therefore run a large suite of models sampling from distributions of the key parameters that govern thermal evolution and resulting atmospheric mass: the planetary volatile budget, the atmospheric loss rate, the reference viscosity of the mantle, the radioactive heating budget of the mantle, and the initial mantle temperature. We track which models do or do not successfully result in atmospheres <10 bar in size. As the uncertainty range on the age of LHS 3844 is 6.2–9.4 Gyr, we determine for each model whether it meets the constraints on atmosphere size by both the upper and lower age bounds. The models assume LHS 3844b lies in a stagnant lid regime of tectonics. While the tectonic mode of any exoplanet is unconstrained, we assume a stagnant lid because the high surface temperature of LHS 3844b and lack of surface water are expected
transition in model behavior, where models with low CO$_2$ budgets always end with a <10 bar atmosphere, while those with high CO$_2$ budgets nearly always end with atmospheres >10 bar. We start our models with all of the planet’s CO$_2$ initially residing in the mantle, and then we calculate how CO$_2$ accumulates in the atmosphere over time by degassing. Some CO$_2$ may be outgassed during a putative magma ocean stage, which is neglected here. Including magma ocean degassing would only serve to hasten the rapid accumulation of mantle C into the atmosphere that the large majority of our models predict.

The models track the evolution of atmosphere size assuming that volcanic degassing of CO$_2$, as calculated from the thermal evolution model, determines the rate at which mass is added to the atmosphere, and that the rate at which atmosphere is lost is determined by a constant stripping rate. The erosion of planetary atmospheres is an area of growing research efforts, particularly in the environment of low-mass stars where periods of sustained stellar activity can occur (Lammer et al. 2008; Rodríguez-Mozos & Moya 2019). Based on the stellar activity discussion in Section 2.2, we adopt the range of atmospheric loss rates scaled by the loss rate from Proxima Cen b of between 30 and 300 kg s$^{-1}$ (Dong et al. 2017). The atmospheric loss rate in a particular model is held fixed in time. As atmospheric erosion cannot occur if no atmosphere is present, we scale the atmospheric loss rate linearly with atmosphere size when the atmosphere mass is <44,000 kg (or <10$^5$ moles of CO$_2$), above this threshold, atmospheric loss rate is constant.

We also assume a mantle reference viscosity (the viscosity of the mantle at the present day interior temperature of the Earth) of 10$^{18}$-10$^{22}$ Pa s, bracketing the range of typical estimates for the Earth. The radioactive heat budget of the planet has an important control on thermal evolution and outgassing (Foley & Smye 2018). We consider a range of 50%-200% of the Earth’s heat-producing element budget, based on the observational bounds of radionuclides from stellar abundances (Unterborn et al. 2015; Botelho et al. 2019). Finally, we also consider a range of initial mantle potential temperatures of 1700-2000 K. For each of the three end-member interior structures, one million models are run (three million models in total). The models sample from uniform distributions of total C budget, atmospheric loss rate, mantle reference viscosity, heat-producing element budget, and initial mantle temperature. Logarithmic distributions are used for C budget, loss rate, and reference viscosity.

5. Model Results

Most models show rapid degassing of CO$_2$ within ~1 Gyr (Figure 4). During this early outgassing stage, the rate of outgassing far exceeds the atmospheric loss rate, so atmospheric size rapidly increases. Typically, nearly all of the planet’s supply of CO$_2$ is outgassed during this time, after which growth of the atmosphere essentially stops; atmospheric stripping then dominates and the size of the atmosphere shrinks for the rest of the planet’s lifetime. Rapid early outgassing is a result of extensive early volcanism, which is expected for stagnant lid planets due to high initial rates of internal heat production and primordial heat. As LHS 3844b almost certainly lacks surface water, CO$_2$ cannot be removed from the atmosphere by weathering and the formation of carbonate rocks. This means there is no mechanism for returning C from the atmosphere to the mantle, and outgassing irreversibly depletes the mantle of C. Once the supply of C in the mantle has been depleted, degassing rates become negligible, even as volcanism continues, and atmospheric growth stops. In the large majority of our models, mantle volcanism continues for more than ~1 Gyr, typically lasting ~3-5 Gyr.

As most models outgas nearly all of their CO$_2$ to the atmosphere (as in the examples shown in Figure 4), the size of atmosphere formed is largely controlled by the mantle C budget. With $C_{\text{conc}} \sim 10^{-4}$ wt% (two orders of magnitude smaller than Earth’s C budget in terms of mantle wt%), the peak atmosphere size that forms is ~2 bar; with $C_{\text{conc}} \sim 10^{-3}$ wt% (one order of magnitude smaller than Earth), peak atmosphere size is ~20 bar; and with $C_{\text{conc}} \sim 10^{-2}$ wt% (approximately equal to Earth’s C budget), it is ~200 bar. High atmospheric stripping rates of 300 kg s$^{-1}$ can remove enough atmosphere to bring a ~20 bar atmosphere that forms early in the planet’s history down to <10 bar after ~8 Gyr. Significantly higher loss rates would be required to remove the ~200 bar of atmosphere that forms at the upper end of mantle C budgets we explore. Geophysical factors, such as heat-producing element budget, mantle viscosity, and initial mantle temperature, do not have a significant influence on the size of atmosphere formed or size of atmosphere remaining at LHS 3844b’s current age. Lowering initial mantle temperature or increasing the reference viscosity can delay the onset of volcanism, leading to different histories in the first few hundred million years of evolution, but atmosphere size converges in these models at later times (Figure 4(B) and (C)). However, there are combinations of radionuclide budget, mantle reference viscosity, and initial temperature that can significantly limit—and even entirely prevent—mantle degassing, as discussed later in this section.

Our estimates of atmosphere size are generally consistent with Dorn et al. (2018). Dorn et al. (2018) estimate an atmosphere of ~70 bar for a stagnant lid planet of similar mass to LHS 3844b. Their models assume 1000 ppm as the concentration of CO$_2$ in the mantle, which corresponds to ~10$^{-3}$ wt%, the middle of the range of CO$_2$ concentrations we consider; with this concentration, we estimate a comparable CO$_2$ atmosphere size of ~20 bar. Dorn et al. (2018) also keep the concentration of CO$_2$ in the mantle fixed over time, rather than treating it as a CO$_2$ reservoir that shrinks due to outgassing, though they do track mantle depletion and assume depleted regions have their volatile abundances lowered by melting. This difference may explain the larger CO$_2$ atmosphere they estimate (~70 bar versus ~20 bar).

The examples shown in Figure 4 demonstrate that atmospheric loss rate and mantle C budget have significant control over the ending atmosphere size. To elucidate the full combination of model parameters that result in atmospheres fitting the observational constraints for LHS 3844b, we show histograms of all successful models in Figure 5. There is a clear trend of low C budgets and high loss rates being able to explain the thin atmosphere. For all three assumed interior structures we modeled, $C_{\text{conc}} < 5 \times 10^{-4}$ wt% results in a large number of successful models; the number of successful models then decreases rapidly from $C_{\text{conc}} \approx 5 \times 10^{-4} - 10^{-3}$ wt%, and very few successful models result from $C_{\text{conc}} > 10^{-3}$ wt%. Thus models with C budgets of approximately an order of magnitude lower than Earth’s or larger almost always produce atmospheres larger than the observational constraint.

Successful models are found for all modeled atmospheric loss rates, but higher loss rates clearly lead to a higher proportion of successful models. Successful models are found
in near-uniform distributions of heat-producing element budget, mantle reference viscosity, and initial mantle potential temperature. There is a small increase in the proportion of successful models with low heat-producing element budgets. This indicates that producing an atmosphere <10 bar is not sensitive to initial mantle temperature or reference viscosity, and largely insensitive to radionuclide budget, save for slightly favoring low radionuclide budgets. An important feature of the results is that the assumed interior structure does not have a significant impact; the Earth-like, small core, and large core cases all show very similar distributions of successful models among the considered parameters. Whether the low or high end of the estimated age range for LHS 3844b is used also does not have a significant effect, though intuitively, more models are successful when the upper bound on age is used.

The influence of mantle C budget and loss rate can be easily explained. First, there is a limit below which there is simply not enough C in the mantle to produce an atmosphere >10 bar. Below this limit, all models will be successful, as it will not be possible to form an atmosphere above the current observational constraints. As the atmospheric pressure is given by

$$P_{\text{atm}} = \frac{C_{\text{atm}} m_{\text{CO}_2} g}{A_s},$$

where $C_{\text{atm}}$ is the number of moles of CO$_2$ in the atmosphere, $m_{\text{CO}_2}$ is the molar mass of CO$_2$, and $A_s$ is the surface area of the planet, then for

$$C_{\text{tot}} < \frac{(10 \text{ bar}) A_s}{m_{\text{CO}_2} g},$$

all models will be successful; this limit is found to be $1.25 \times 10^{21}$ mol (C$_{\text{conc}} = 5.14 \times 10^{-4}$ wt%) for an Earth-like core, $1.37 \times 10^{21}$ mol (C$_{\text{conc}} = 5.18 \times 10^{-4}$ wt%) for the small core case, and $1.04 \times 10^{21}$ mol (C$_{\text{conc}} = 4.50 \times 10^{-4}$ wt%) for the large core case. These limits are shown in Figure 5 as vertical dashed lines.

For larger $C_{\text{tot}}$ (or C$_{\text{conc}}$), an atmosphere >10 bar can form if significant outgassing occurs. Thus, atmospheric stripping is required to bring the atmosphere back to within observational constraints. Histograms of all successful models with C$_{\text{conc}} > 5.14 \times 10^{-4}$ wt% for the Earth-like core size case are shown in Figure 6; similar distributions are seen for the small core and large core models (not shown). There is a clear tradeoff between C$_{\text{conc}}$ and loss rate. Higher loss rates allow larger C$_{\text{conc}}$ values to still satisfy the atmosphere size constraint, while lower C$_{\text{conc}}$ values allow even low atmospheric loss rates to still result in successful models. Specifically, for a given loss rate, there is a maximum atmospheric size that can be formed early in the planet’s history and still be stripped to <10 bar after 6.2–9.4 Gyr. This limit on maximum atmospheric size is illustrated in Figure 6(D), which shows the peak atmospheric size reached during each successful model run as a function of loss rate. For a

Figure 4. Time evolution of total atmospheric pressure, $P_{\text{atm}}$, for example models with an Earth-like core size. Variations from a baseline model with an Earth-like heat-producing element (HPE) budget, mantle CO$_2$ concentration of $C_{\text{conc}} = 1.2 \times 10^{-3}$ wt%, initial mantle temperature $T_{\text{init}} = 2000$ K, and reference viscosity $\mu_{\text{ref}} = 1.3 \times 10^{20}$ Pa·s are shown: (A) variations in HPE budget; (B) variations in mantle reference viscosity; (C) variations in initial mantle temperature; and (D) variations in mantle C concentration. For all figures, solid lines denote an atmospheric loss rate of 30 kg s$^{-1}$, and dashed lines a loss rate of 300 kg s$^{-1}$. Dotted line shows 10 bar upper bound on LHS 3844b’s atmosphere size based on observational constraints.
loss rate of 300 kg s\(^{-1}\), atmospheres can only reach \(\approx 20\)–25 bar and still be stripped to <10 bar within LHS 3844b’s lifetime. Our assumed upper limit on atmospheric loss rate is simply not high enough to remove atmospheres larger than 20–25 bar.

In fact, we can still make a simple estimate of this maximum atmospheric pressure \(P_{\text{atm}}^{\text{max}}\) that can be reached and still satisfy the constraint on present-day atmosphere size. Assuming that the atmosphere forms very early in the planet’s history, as seen in our outgassing models, the total mass of atmosphere that can be stripped is given by the product of the loss rate \(E\) and the planet age \(\tau\); thus

\[
P_{\text{atm}}^{\text{max}} = 10\text{ bar} + \frac{E\tau}{A_2},
\]

for \(E = 300\) kg s\(^{-1}\), \(\tau = 9.4\) Gyr, and \(g = 15.7\) m s\(^{-2}\), as for an Earth-like core size, \(P_{\text{atm}}^{\text{max}} \approx 25\) bar, nearly identical to the upper limit seen in Figure 6(D). Further assuming that all of the planet’s mantle CO\(_2\) budget is outgassed, the upper bound on mantle C budget that can explain the present-day state of LHS 3844b’s atmosphere, as a function of atmospheric loss rate, can be calculated as

\[
C_{\text{tot}}^{\text{max}} = \frac{10\text{ bar} \times A_1}{m_{\text{CO}_2} g} + \frac{E\tau}{m_{\text{CO}_2}}.
\]

With the same numbers as listed above, \(C_{\text{tot}}^{\text{max}} \approx 3 \times 10^{21}\) mol, which corresponds to a mantle C concentration of about 10% Earth’s. For LHS 3844b to have an Earth-like mantle C budget, the atmospheric stripping rate would have to be a factor of 10 larger than our estimated upper bound.

Finally, Figure 6(C) shows that more models are successful when the heat-producing element budget is low. This reveals one additional group of models that are able to produce atmospheres of <10 bar: models with very limited (or even completely absent) volcanic outgassing. To illustrate the factors that allow planets to experience limited outgassing, we plot all models that produce maximum atmospheric pressures <10\(^{-6}\) Pa in Figure 7. Here, we show only the Earth-like core and large core models, as very few of the small core models resulted in such limited outgassing. The cutoff of 10\(^{-6}\) Pa was used because there was a clear grouping of models with maximum atmospheric sizes below this threshold when looking at the entire suite of models together. Models that experience essentially no outgassing are characterized by low heat-producing element budgets, low initial mantle potential temperatures, and high mantle reference viscosities. All of these factors act to suppress mantle melting. Low initial temperatures and radionuclide budgets limit mantle heating and can keep the mantle cool enough not to melt and produce volcanism. A high reference viscosity increases the thickness of the stagnant lid, which also suppresses volcanism. A thicker stagnant lid requires a higher mantle temperature for melting to occur, because it forces melting to occur at higher pressures. In the large core models, there is also a trend toward low C budgets and high loss rates, as these factors can keep the atmospheric size low even if some limited volcanism occurs.

![Figure 5](https://example.com/figure5.png)

**Figure 5.** Histograms of all model results that successfully produce an atmosphere of <10 bar after 6.2 Gyr (blue) and 9.4 Gyr (red). Top row shows models with an Earth-like core size, middle row models with a small core, and bottom row models with a large core. Shown are the number of successful models as a function of the mantle C budget (A, F, and K), the atmospheric loss rate (B, G, and L), the heat-producing element budget relative to Earth’s (C, H, and M), the mantle reference viscosity (D, I, and N), and the initial mantle temperature (E, J, and O). Dashed lines show the limit where the mantle C budget is low enough that even complete outgassing of all mantle C will result in a <10 bar atmosphere (A, F, and K). Plots of mantle C budget are presented in terms of mantle C concentration, \(C_{\text{conc}}\) in wt% (lower axis) and mantle C concentration normalized to Earth, \(C_{\text{conc}}/C_{\text{conc,E}}\) (upper axis).
However, it should be noted that volcanism can be shut off by purely geophysical factors—mantle heat budget and viscosity—in which case no atmosphere will form even if the C budget is very large or loss rate is small.

5.1. Carbon in the Core

The thermal evolution models indicate that a mantle depleted in carbon relative to the Earth is the most likely explanation for the present-day thin atmosphere of LHS 3844b. However, whether a C-depleted mantle means the overall planet composition is depleted in carbon relative to the Earth depends on the partitioning of carbon between the mantle and core. For a given bulk planet composition of carbon, the core composition and initial mantle concentration are set as the core segregates during the planet’s magma ocean stage (Dasgupta et al. 2013; Fischer et al. 2020). Geochemical estimates, based on meteoritic abundances, show the bulk Earth (crust + mantle + core) contains 0.07 wt% carbon total (McDonough 2003).

Of this bulk carbon, the Earth’s core contains ~90% of the total budget, yielding a core C concentration of 0.2 wt%. This leaves only 0.01 wt% C in the mantle (McDonough 2003). This ratio of the mass of C split between the mantle and core is known as the partition coefficient \(D_C = f_{C,\text{mantle}} / f_{C,\text{mantle}}\). These geochemical models predict \(D_C = 9\) for the Earth. However, \(D_C\) is a complex function of the pressure, temperature, and oxygen fugacity \((f_{O_2})\) of the magma ocean upon equilibration, the fraction of sulfur and oxygen in the iron melt, and the degree of silicate melt polymerization (Fischer et al. 2020). We define the affinity for an element to partition into metallic iron as siderophile. Generally, a larger fraction of C enters the mantle—that is, C becomes less siderophile—as pressure, degree of melt polymerization, and the fraction of S in the melt increase. Conversely, more C enters the core, that is C becomes more siderophile, as the fraction of O in the melt and \(f_{O_2}\) increase. Core segregation can happen in two ways: as a single event where material equilibrates at pressures near the core–mantle
boundary; and in a multistage fashion where accreting material slowly adds volatiles to both the core and mantle, leading to much lower equilibration pressures.

We model the competing effects of pressure, composition, and oxygen fugacity on C partitioning into the core assuming single-stage core formation. We use the formulation of Fischer et al. (2020) for determining $D_C$ and thus estimate the fraction of C that is retained in the mantle for a given bulk C budget of LHS 3844b. We assume a degree of polymerization comparable to peridotite (2.6), and in order to maximize the fraction of C in the core, set the fraction of S in the melt to zero. We then construct a model of $D_C$ containing four variables: pressure ($P$), temperature ($T$), the fraction of O in the metal ($X^{\text{metal}}_O$), and $fO_2$ of the material when the metal and silicate equilibrate. The value of $X^{\text{metal}}_O$ is also a function of $P$, $T$, and $fO_2$. Using the exchange coefficient ($K_D$) formulation as a function of $P$ and $T$ of Fischer et al. (2015) for O, we can estimate the fraction of O in the melt using the relation: 

$$K_D = (X^{\text{metal}}_{\text{Fe}} - X^{\text{metal}}_O)/(X^{\text{silicate}}_\text{Fe} - X^{\text{silicate}}_O).$$

The equilibrium ratio $X^{\text{metal}}_{\text{Fe}} / X^{\text{silicate}}_{\text{Fe}}$ is determined using the definition of the iron-wustite fugacity buffer that assumes the activity of Fe and FeO in the system are approximated by their mole fractions:

$$\Delta IW = 2 \log_{10} \left( \frac{X^{\text{silicate}}_{\text{Fe}}}{X^{\text{metal}}_{\text{Fe}}} \right).$$

We adopt the liquidus of Stixrude & Lithgow-Bertelloni (2011) to calculate the minimum temperature of the magma ocean as a function of pressure.

The value of $X^{\text{metal}}_O$ increases with increasing equilibration pressure (Figure 8, top). We note that the highest pressure for measurements in Fischer et al. (2015) is $\sim 100$ GPa at $\Delta IW = -1.1$, which estimates $X^{\text{metal}}_O$ of 0.27. For $fO_2 > \Delta IW - 2$ and high pressure, $X^{\text{metal}}_O$ rapidly increases, likely due to our assumption that $X^{\text{silicate}}_{\text{Fe}} / X^{\text{metal}}_{\text{Fe}}$ approximates their activities at high pressure and temperature. Additionally, our simple model only assumes equilibration in the Fe–C–O system. Importantly missing from this model is the partitioning of Si between the mantle and core during formation. The relative fractions of Si and O entering the core are inversely correlated (Fischer et al. 2015), with relatively more Si entering the core under reducing conditions. As such, we likely overestimate $X^{\text{metal}}_O$ and $D_C$ at high pressures and temperatures, particularly under oxidizing conditions where our model predicts $>80\%$ of the mass of metal will be O. We therefore set an arbitrary maximum value of $X^{\text{metal}}_O = 0.45$ and will explore the consequences of relaxing this assumption below.

From $X^{\text{metal}}_O$, we can derive the partition coefficient of C, $D_C$, as a function of pressure and the $fO_2$ of the magma ocean (Figure 8, bottom). Fischer et al. (2020) provides two determinations of $D_C$ produced using either fits to nanoSIMS- or electron microprobe-derived data sets, with the $D_C$ values from the microprobe fit being higher than the nanoSIMS fit (Figure 8, bottom). For oxidizing magma ocean conditions ($fO_2 > \Delta IW - 2$), $D_C$ increases until between 150 and 200 GPa, where it begins to drop as the effect of increasing pressure lowering $D_C$ begins to dominate due to our arbitrary maximum value of $X^{\text{metal}}_O$ limiting the amount of O (and thus C) available to partition into the core. Under reducing conditions, $X^{\text{metal}}_O$ remains low and $D_C$ decreases over all pressures.

Under single-stage core formation, equilibration occurs at roughly 35\% of the central pressure of the planet (Schaefer et al. 2017). For LHS-3844b, we estimate this occurs at $\sim 350$ GPa using the pressure estimates of a 1.3 $R_\oplus$ planet from Unterborn & Panero (2019). Adopting the average $D_C$ of the microprobe and nanoSIMS fits, we estimate that if LHS 3844b formed with an Earth-like concentration of C (0.07 wt\%$) and an Earth-like CMF (0.33), then under oxidizing conditions ($\Delta IW > -2$), little C is sequestered into the core, producing a mantle with 0.03 wt\% C, a factor of three greater than the Earth’s (Figure 9, left). Interestingly, if core equilibration happened between 80 and 280 GPa under these oxidizing conditions, significant amounts of C would partition into the core, leaving behind a mantle significantly depleted in C, akin to what has been found for Earth-size planets (Li et al. 2015, 2016).
Under reducing conditions ($\Delta IW < -2$), $D_C$ is low and practically no carbon is partitioned into the core, producing a mantle with an order-of-magnitude greater C concentration than the Earth. Exploring the effects of bulk C budget, we find that if LHS 3844b formed with 1% of the Earth’s C budget (0.0007 wt%), regardless of the oxidation state of the magma ocean, the mantle will have a concentration below that of the Earth, whereas for a planet with four times the C budget of Earth (0.28 wt%), the mantle will have an even greater C concentration than our Earth concentration model does (Figure 9, center, left). These relatively simple models show that, under single-stage core formation, the mantle’s C concentration is primarily a function of the total amount of C present in the planet. In order to produce a mantle with a C budget low enough to result in a $a \leq 10$ bar atmosphere for LHS 3844b (0.002 wt%, Figure 5), the planet overall would need to be depleted in C relative to the Earth.

Under single-stage core formation, little C enters LHS 3844b’s core, regardless of magma ocean oxidation state. For those models where bulk is at or higher than Earth’s, this lack of C partitioning into the core will produce a magma ocean at carbon saturation. When a magma ocean is at saturation, some fraction of the C will dissolve into the silicate melt, with the rest likely becoming outgassed. Under reducing conditions ($fO_2 < IW - 2$), the solubility of C in a silicate melt at the low pressures ($P \sim 3$ GPa) where degassing occurs is $\sim 150$ ppm (Grewal et al. 2020), as indicated by the gray dashed line in Figure 9. Our reduced Earth-composition model of LHS 3844b predicts a mantle concentration of $\sim 1000$ ppm (Figure 9, left), whereas the low C-content model will retain all of its C in the mantle, likely as reduced C in the form of diamond or graphite (Unterborn et al. 2014). Under oxidizing conditions ($fO_2 > IW - 2$), C solubility in silicate melt increases to $\sim 250$ ppm at $fO_2 = \Delta IW - 2$ and 450 ppm for $fO_2 = \Delta IW - 1.5$ at these pressures (Figure 9, gray box). In this case, only the model with four times the bulk C budget of the Earth will produce a mantle above this solubility range. In both the oxidizing and reducing cases, for those mantle contents above their respective solubilities, mantle degassing of C-bearing gas will occur until mantle C contents become comparable to the solubility. These degassed proto-atmospheres will be comprised of CO2 and CH4 for magma oceans with oxidizing and reducing $fO_2$, respectively (Kasting et al. 1993; Gaillard & Scaillet 2009; Grewal et al. 2020). Even after the considerable degassing, the final mantle composition will still remain above the Earth’s (Figure 9, thick black line).

Under single-stage core formation, the high pressure of equilibration prevents C from partitioning into the core. This means a planet’s bulk C budget coupled with the solubility of C in the silicate melt effectively sets the initial mantle C budget and whether or not a proto-atmosphere will be created. Notably, this proto-atmosphere must also be eroded to explain LHS 3844b’s current lack of atmosphere. Under reduced core formation scenarios, $D_C$ decreases with increasing pressure, meaning C-rich mantles comprised of silicates with graphite/diamonds and a CH4 atmosphere are the likely initial state of rocky super-Earths after magma ocean crystallization (Grewal et al. 2020). Under oxidizing conditions, however, only if $X_{\text{metal}}^{\text{metal}}$ is greater than our arbitrary value of 0.45 will $D_C$ increase at high pressure, allowing for more C to partition into the core. No experiments in the Si–O–Fe–C system have been performed above $\sim 60$ GPa, and thus we are extrapolating well beyond the range of experimental data. Regardless, to increase $X_{\text{metal}}^{\text{metal}}$ values above 200 GPa, $K_D$ must also increase. Otherwise, under single-stage core formation, C is simply not siderophile enough at oxidizing fugacities to partition significant amounts of C into the core. We note, however that the primary oxidant in this case would be water. Much like CO2 or CH4, the amounts of water needed to produce these oxidizing $fO_2$ will also produce significant water vapor and surface oceans (Elkins-Tanton 2011), which must also be eroded in
addition to the CO$_2$ to explain LHS 3844b’s current lack of atmosphere.

Multistage core formation, where accreting material equilibrates with the magma ocean, is another viable core formation scenario (Rubie et al. 2011; Fischer et al. 2017, 2020). In multistage core formation, the planet’s core and mantle grow via a series of equilibration processes, as infalling material equilibrates at low pressures initially when the planet is small, and equilibration pressure increases as the planet grows. This allows for core formation to occur at much lower pressures than single-stage core formation, with equilibration pressures reaching a maximum of ~65% of the CMB pressure that would occur if it formed as a single event (Rubie et al. 2011; Fischer et al. 2017). For the Earth, this pressure is ~80 GPa, compared to its CMB pressure of ~125 GPa. For LHS 3844b, 65% $P_{\text{CMB}} \sim 230$ GPa (Figure 9). Under oxidizing conditions, as material is added and the pressure of equilibration increases, O and C become more siderophile. As such, the C of early infalling material will equilibrate at low pressures and its C will stay in the mantle. Later infalling material, however, will equilibrate at higher pressures and the majority of the accreting material’s complement of C will partition into the core. This scenario is only likely, though, if LHS 3844b formed entirely out of oxidized material. The Earth, however, likely formed from a mixture of oxidizing and reduced material, which would explain its trace element abundances (Wade & Wood 2005). If LHS 3844b also formed from a mixture of material of varying $f_{O_2}$, the mantle may still be left C-rich due to the low $D_C$ of the reducing material, despite some fraction of the planet’s C being partitioned into the core as a consequence of accreting oxidizing material as well. Whether the balance of the low pressure of equilibration in multistage formation and the planet forming from a mixture of reduced and oxidizing material are able to produce C-poor mantles for planet’s with Earth-like or greater C budgets is beyond the scope of this paper. In the absence of new experimental data for the partitioning of O, Si, and C at $P > 150$ GPa, in both single-stage core formation and multistage core formation, our models of the carbon content of LHS3844b’s mantle effectively reflect that of the bulk composition the planet overall inherits during formation (Figure 9). Therefore, we find that, in order to produce the C-poor mantle the degassing models require, and to avoid a magma ocean-derived proto-atmosphere that would also need to be eroded away in addition to the later degassed atmosphere, LHS 3844b must form volatile-poor compared to the Earth, in both single- or multistage core formation scenarios.

5.2. Can Impacts Strip LHS 3844b’s Atmosphere?

Infalling material onto a planet either during accretion (Zahnle et al. 1990, 1992; Svetsov 2007; Shuvalov 2009; Schlichting et al. 2015), or through a giant impact (Newman et al. 1999; Genda & Abe 2003, 2005; Schlichting et al. 2015) can potentially remove some or all of an exoplanet’s atmosphere. We model the amount of material needed to strip an atmosphere using the formalism of Schlichting et al. (2015), which uses analytic self-similar solutions and full numerical integrations to calculate the fraction of an atmosphere of scale height, $H$, lost for an individual impactor of a given radius ($r$) and density ($\rho$), hitting a planet of radius ($R$) with an atmospheric density of $\rho_0$. This formalism yields three key values: the minimum radius of an impactor able to strip any atmosphere ($r_{\text{min}}$), the radius of impactors where it is considered a giant impact that sends strong shocks through the atmosphere ($r_g$), and the number ($N$) of impactors of a given size ($r$) needed to entirely strip the atmosphere.

We apply this formalism to atmospheres composed of two different compositions: an oxidized atmosphere containing only CO$_2$, and a reduced atmosphere containing only CH$_4$. We adopt two accretion scenarios where LHS 3844b begins at 2.2 and 2.7 $M_{\oplus}$ and is accreting the final 0.2 $M_{\oplus}$ of its mass, rendering our end-member predicted masses for the planet (Section 3.1). We also assume this proto-LHS 3844b is at roughly its current radius of 1.303 $R_{\oplus}$, as 0.2 $M_{\oplus}$ of added material will not change the radius in any significant way (Unterborn & Panero 2019). This yields initial gravities for LHS 3844b between 12.7 and 15.6 m s$^{-2}$. The model of Schlichting et al. (2015) calculates the minimum radial size of...
an impactor to strip any quantity of atmosphere, \( r_{\text{min}} \):

\[
r_{\text{min}} = H \left( \frac{3\rho_0}{\rho} \right)^{1/3},
\]

where \( H \) is the scale height of the atmosphere, \( \rho_0 \) is the density of the atmosphere at the surface, and \( \rho \) is the density of the impactor. We set \( \rho \) to 2 g cm\(^{-3} \) below impactor sizes \( r = 1000 \text{ km}, 3 \text{ g cm}^{-3} \) for 1000 \( \leq r < 3000 \text{ km} \) and \( 4 \text{ g cm}^{-3} \) for \( 3000 \leq r < 5000 \text{ km} \) and \( 5.5 \text{ g cm}^{-3} \) for impactors of size \( r \geq 5000 \text{ km} \). To determine scale height, \( H \), we assume an atmospheric temperature of 1000 K and atmospheric pressure of 30 bar, then use the ideal gas law to determine the density of the atmosphere \((\rho_0)\) and our gravities derived above. We define \( r_{\text{gi}} \) as:

\[
r_{\text{gi}} = (2HR^2)^{1/2},
\]

where \( R \) is the radius of the planet \((1.303 \text{ } R_\oplus)\). We calculate values of \( r_{\text{gi}} \) between \( \sim 1200 \text{ km} \) for a pure CO\(_2\) atmosphere and 1700 km for a pure CH\(_4\) atmosphere. To quantify the mass of atmosphere ejected during to an impact relative to the mass of the impactor for small impact angles, \( \frac{M_{\text{ejected}}}{M_{\text{impactor}}} \), we use Equation (39) of Schlichting et al. (2015):

\[
\frac{M_{\text{ejected}}}{M_{\text{impactor}}} \leq \frac{r_{\text{min}}}{2r} \left( 1 - \left( \frac{r_{\text{min}}}{r} \right)^2 \right).
\]

where \( r \) is the radius of the impactor. We find that \( M_{\text{ejected}}/M_{\text{impactor}} \) increases with impactor radius, reaches a maximum at \( \sim 1 \text{ km} \), and decreases afterward (Figure 10). Additionally, CH\(_4\) atmospheres are much easier to strip than CO\(_2\) atmospheres are, due to their factor of \( \sim 4 \) higher scale height. Furthermore, impactors with radii of 200 km have \( M_{\text{ejected}}/M_{\text{impactor}} \) an order of magnitude higher than impactors meeting the threshold for giant impacts (Figure 10). While a single giant impact will strip some fraction of an atmosphere, atmospheric loss by accretion of many small object is a much more efficient process per unit mass, comparatively (Schlichting et al. 2015).

The total number \((N)\) of impactors of a given size needed to completely strip an atmosphere is a function of the size of the atmosphere and the amount of atmosphere ejected due to an impact of a given size (Schlichting et al. 2015). To simplify this model, we assume that these impactors are not carrying additional volatiles to be added to the atmosphere. Using Equations (47), (49), and (51) of Schlichting et al. (2015), which assume the impactor is traveling at roughly the escape velocity of the planet, we find that, for a CO\(_2\) atmosphere, roughly \( N \sim 10^{3-8} \) small impactors \((r < r_{\text{gi}})\) are needed to entirely strip the atmosphere, with the exact value depending on the impactor’s size (Figure 11, top). We find slightly smaller values of \( N \) for CH\(_4\) atmospheres, likely due to the larger scale height compared to CO\(_2\). Above \( r_{\text{gi}} \) for both species, \( N \) slowly drops as impactor radius increases, reaching \( \sim 30 \) accretionary impacts needed for Mars-sized impactors and \( \sim 3 \) needed for Earth-sized impactors.

While accretion can strip an atmosphere, it also adds to the overall mass of a planet. Our initial masses require only 0.2 \( M_\oplus \) of material to reach our final estimated masses of LHS 3844b \((2.4 \text{ and } 2.9 \text{ } M_\oplus)\). For those impactors where the total amount of accreted material exceeds 0.2 \( M_\oplus \), some fraction of the protoplanet must also be stripped to space upon impact. For protoplanets that have undergone core segregation, the mantle will be stripped, leaving behind a planet with a relatively large CMF. For a CO\(_2\) atmosphere, the size of impactor where mantle stripping becomes necessary is \( \sim 450 \text{ km} \) for both our small and large initial planet masses (Figure 12, solid). For a CH\(_4\) atmosphere, these values rise to \( \sim 900 \text{ km} \) (Figure 12, dashed). These differences due to atmospheric composition are purely a function of the factor of \( \sim 4 \) difference in the scale
impactors are considered giant for atmospheres of CO₂ are shown as red and blue dotted 3844b if it had a relatively small core (2.4 \( M_\oplus \)) or relatively large core (2.9 \( M_\oplus \)) are shown as red and blue dotted–dashed lines, respectively. Radii above which impactors are considered giant for atmospheres of CO₂ (solid) and CH₄ (dashed) are labeled as black lines.

The heights of these gasses. Many more than \( N \) impactors of these sizes can accrete; however, mantle stripping would become required in such a case, in order to reduce the overall mass. This does not mean that impactors below this size cannot strip mantle, but rather that, if all infalling material is above this size, mantle stripping must occur in order to match our mass estimates of LHS 3844b. For all of our models, giant impact scenarios require significant mantle stripping to match LHS 3844b’s mass regardless of atmosphere composition.

These previous models assume that the impactors velocity is roughly that of the escape velocity. However, as the velocity of impact changes, so does the relative amount of atmosphere ejected. For example, both Earth and LHS 3844b require \(~30\) impactors of Mars size (\(~3400\) km) to strip their atmospheres completely when traveling at the escape velocity (\( v_\text{escape} \)). The current orbital velocity of LHS 3844b is \(~150\) km/s. Using Equation (32) from Schlichting et al. (2015), for a 500 km object to remove a \( 1/N \) amount of atmosphere (\( N = 300 \)) from a 2 \( M_\oplus \) LHS 3844b, it must travel \(~70\) times the escape velocity for both a pure CO₂ and CH₄ atmosphere. For a Mars-sized object (\( r = 3400 \) km, \( N = 30 \)), this velocity is reduced to essentially the escape velocity. If accreting material impacts at speeds slower than these, \( N \) must increase in order to fully strip the atmosphere.

While \( N \) is a measure of how many accreting impacts traveling at roughly \( v_\text{escape} \) are needed to strip an atmosphere, we can also estimate the impact velocity (\( v_\text{impact} \)) needed to entirely strip the atmosphere in a single giant impact after LHS 3844b has fully formed with a radius of 1.303 \( R_\oplus \) and between 2.4 and 2.9 \( M_\oplus \). Using Equation (32) of Schlichting et al. (2015) for an isothermal atmosphere and setting the fraction of atmosphere loss to 1, we estimate the relative velocity of an impactor to the planet’s escape velocity (\( v_\text{impact} / v_\text{esc} \)) needed to completely strip the atmosphere in one impact as a function of the impactor’s mass relative to LHS 3844b’s (\( m_\text{impact} / M_{\text{LHS} 3844b} \)). We find that, as \( m_\text{impact} / M_{\text{LHS} 3844b} \), the \( v_\text{impact} / v_\text{esc} \) needed to eject an atmosphere entirely decreases (Figure 13). The value of \( v_\text{esc} \) varies with the initial mass of the planet, between 10.2 and 11.4 km/s for our range of initial mass models. For a Mars-sized object, it must travel \(~100\) times the escape velocity, whereas if this Mars-sized object was to hit the Earth, it would be required to travel at only \(~10\) times the escape velocity (\( m_\text{impact} / m_\oplus = 0.1 \)). For an Earth-sized impactor hitting LHS 3844b, it need only travel at \(~1.3–1.4\) times the escape velocity to strip the atmosphere entirely. These velocities are roughly 70 times the orbital velocity for a Mars-size object and only 10% for an Earth-size object. We calculate the minimum size of an object that can strip an atmosphere entirely in one impact while traveling at the escape velocity to be 11%–12% of the mass of LHS 3844b, or 0.26–0.35 \( M_\oplus \). Numerical results have found that multiple giant impacts like these are likely in models of planets Earth-size and slightly larger (\( M < 1.6 M_\oplus \)) in the Solar protoplanetary disk (Quintana et al. 2016), with Kepler-107c showing evidence that it underwent a mantle-stripping giant impact (Bonomo et al. 2019). Quintana et al. (2016) also found, however, that the planets in their model only experience an average of three giant impacts (up to a maximum of eight in their models), all of which were unlikely to fully strip its atmosphere or oceans. Whether this is also true for systems like LHS 3844b is not known. However, if the planet was impacted by a single Earth-size impactor traveling at or below the orbital velocity of LHS 3844b, complete atmospheric ejection and significant mantle stripping is likely. Comparatively, a single impact from a Mars-sized object is a less plausible scenario for stripping the atmosphere of LHS 3844b, due to the high impact velocity needed.

![Figure 12](image1.png)

Figure 12. Final mass of LHS 3844b starting at 2.2 \( M_\oplus \) (red) and 2.7 \( M_\oplus \) (blue) after accreting the \( N \) impactors of a given radius needed to strip its CO₂ (solid) or CH₄ (dashed) atmosphere. Our final estimates for the mass of LHS 3844b if it had a relatively small core (2.4 \( M_\oplus \)) or relatively large core (2.9 \( M_\oplus \)) are shown as red and blue dotted–dashed lines, respectively. Radii above which impactors are considered giant for atmospheres of CO₂ (solid) and CH₄ (dashed) are labeled as black lines.

![Figure 13](image2.png)

Figure 13. Relative velocity of impactor to the escape velocity of LHS 3844b needed to entirely strip an atmosphere with a single impact as a function of the mass ratio of the impactor to the proto-LHS 3844b for initial planet masses of 1 \( M_\oplus \) (blue), 2 \( M_\oplus \) (red), and 2.7 \( M_\oplus \) (black). Ranges of relative masses for Mars (orange) and Earth (green) across our estimated masses of LHS 3844b are shown for reference.
Either through the accretion of small objects \((r < r_g)\) or single giant impact events, atmospheric stripping is only effective if there is an atmosphere to eject. Degassing from a planet’s interior is not an instantaneous process. If this final 0.2 \(M_⊕\) of mass accretes onto LHS 3844b prior to degassing from the magma ocean or subsequent degassing from the solid mantle, the material will simply accrete as normal and no atmospheric stripping will occur. Elkins-Tanton (2011) found that significant volatile degassing due to the solidification of the magma ocean happens over 5 Myr. Pebble accretion tends to form planets via accretion relatively rapidly, on the order of <1 Myr (Johansen & Lambrichts 2017), making it unlikely that accretionary impacts during the magma ocean phase will have any atmosphere to strip. Unless material is delivered later, after the magma ocean phase, will impacts by smaller objects be a viable scenario for stripping LHS 3844b’s atmosphere. However, in the case of impact stripping after the magma ocean stage, accreting material would have to arrive \(>\sim 100\) Myr after magma ocean solidification for any significant atmosphere to exist, based on our mantle outgassing models. Something akin to a late heavy bombardment would be required in order to strip the out-gassed atmosphere after magma ocean solidification, either with smaller impactors or giant impactors. As impactors with mass above 0.26–0.35 \(M_e\), need only travel at or below the orbital velocity of LHS 3844b to entirely strip the atmosphere, giant impactors can not be ruled out as an explanation for LHS 3844b’s lack of atmosphere. In the case of these giant impacts, however, significant mantle stripping is likely. As such, upon measurement of LHS 3844b’s mass and host-star composition, if LHS 3844b should have a density higher than characteristic of a nominally rocky planet with the host star being comparatively Fe-poor, this would imply that the planet underwent a significantly massive, mantle-stripping, giant impact event, akin to those expected for observed super-Mercuries (Bonomo et al. 2019).

6. Discussion

Our models pose three general possibilities for the composition and dynamical history of LHS 3844b: (1) a giant impact ejected all of its atmosphere; (2) it did not undergo degassing at all, due to its geophysical state; and/or (3) it formed C-poor relative to the Earth (Figure 5, panels (A), (F), (K)). Interestingly, for planets where volcanism and degassing are suppressed entirely, planets larger than Earth core sizes also favor low-C abundances, with a slight favoring of low C for Earth-like core sizes (Figure 7). As discussed in Section 5.2, we cannot definitively rule out LHS 3844b’s atmosphere being ejected by a giant impact, but note that this scenario will likely leave the planet as an Fe-enriched “super Mercury.” If LHS 3844b is found to have a high density upon measurement of its mass, this will lend credence to this scenario explaining the planet’s lack of atmosphere.

Scenario 2 requires that a planet forms cold with an initial mantle potential temperature \(<1800\) K, that it has an interior composed of high-viscosity minerals, or that it forms with a radiogenic heat budget between 50% and 60% of the Earth’s. Most likely, a planet would have to have all three of these aspects to prevent melting. Any planet that formed cold with high viscosity, but with an Earth-like radiogenic heat budget, would simply heat up as energy is released into the mantle from radioactive decay, raising the potential temperature and triggering melting. Similar arguments can be made for each of the other combinations. A test for scenario 1 would be to measure the radionuclide abundances of the host-star LHS 3844, which provide observational constraints for the radiogenic heat budget of the planet (Unterborn et al. 2015).

Scenario 3 argues that sub-Earth concentrations of C are the most likely explanation for the present thin, \(<10\) bar atmosphere. This is due to the low partition coefficient of C limiting how much can be stored in the planet’s core, thus producing a mantle with a C concentration nearly that of the bulk planet’s. For bulk planet compositions at or above the Earth’s, these mantle concentrations are well above the smallest that we predict would produce a \(<10\) bar atmosphere today (0.002 wt%; Figure 1). Therefore, despite the relatively high age of the host star (6–9 Gyr), the atmospheric stripping rate would have to be at least a factor of 10 larger than the estimated upper bound for an Earth-like C abundance to result in a present-day atmosphere \(<10\) bar (see Equation (4)).

A relatively volatile-poor composition of LHS 3844b suggests the planet formed interior to its water snow line (or ice line). Had LHS 3844b formed beyond the snow line, it would likely contain significant amounts of water, similar to those inferred for TRAPPIST-1 (Unterborn et al. 2018a, 2018b; Grimm et al. 2018). This water would then be outgassed at the same time as \(CO_2\), contributing even more mass to the atmosphere, as liquid water is not stable on LHS 3844b’s surface. Assuming LHS 3844b did form outside the snow line, a conservative estimate of its primordial water content would be \(~1%) by mass. For the mass range described in Section 3.1, this equates to \(1.4 \sim 1.7 \times 10^{23}\) kg of water. If the mass-loss rate due to atmospheric erosion is the same as our highest assumed value for \(CO_2\) (300 kg s\(^{-1}\)), only \(8.9 \times 10^{19}\) kg of water could be removed from the planet. Therefore, if LHS 3844b had formed beyond the snow line, a thick \(H_2O\) (or possibly \(O_2\), due to hydrogen escape) atmosphere would still remain, unless the escape rate was \(~4) orders of magnitude larger than our estimates. Interestingly, scenario 2 also implies a relatively water-poor composition of LHS 3844b. This is due to the fact that, as water is added to mantle rocks, it lowers both their melting temperature (Hirschmann 2006) and viscosity (Karato 2011). This allows for melting and degassing to occur at lower temperatures and shallower depths than those of the dry mantle case modeled here. C-rich ices condense at much lower temperatures (and greater orbital distances) than water in disks (Lodders 2003). Any planet forming further out in the disk where C-rich ices are stable would also form in a region where condensed water is also stable. A water- or carbon-rich origin of LHS 3844b is unlikely in both scenarios 2 and 3.

If LHS 3844b formed interior to the water snow line, this allows us to constrain the degree to which it migrated as it formed. Unterborn et al. (2018a) calculated the location of the snow line in M-dwarf disks, assuming a passively heated, flared disk (Chiang & Goldreich 1997). From Unterborn et al. (2018a), we estimate the location of the snow line to be located at 0.26 au at 1 Myr (Figure 14, assuming the reported stellar mass of 0.151 \(M_⊙\) and estimating luminosity of LHS 3844 using the stellar evolution models of Baraffe et al. 2002). Unterborn et al. (2018a) calculated the temperature of water condensation as 212 K, which is greater than that of the solar nebula (170 K), owing to the greater surface density of M dwarf disks compared with the solar nebula.
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Figure 14. Calculated orbital distance of the water snow line for LHS 3844b as a function of time for water condensation temperatures of an M-dwarf disk (212 K; red) and the solar nebula (170 K; blue). Current orbital distance of LHS 3844b is shown as a black dashed line.

The location of the water snow line thus sets the maximum orbital distance at which LHS 3844b could have formed. If LHS 3844b formed later than 1 Myr, this maximum distance decreases as the disk cools (Figure 14). As LHS 3844b cannot cross exterior to the snow line, this also sets the rate at which it must migrate. If LHS 3844b formed early (1 Myr), a rapid migration rate followed by slower one is possible in order to remain interior to the snow line. If LHS 3844b formed at 10 Myr, a much slower migration rate and a smaller migration distance are possible (Lykawka & Ito 2013; Izidoro et al. 2014). It is also possible that LHS 3844b formed entirely in place with no migration. This connection between observed composition and disk chemistry clearly provides constraints on the formation dynamics of the planet itself—and can be applied to other systems.

The volatile-poor nature of LHS 3844b provides additional clues to its formation history. Kreidberg et al. (2019a) place an upper limit for LHS 3844b’s surface water content of <0.02 wt%. This is very similar to the Earth’s concentration of <0.01 wt% water, which is partitioned between the surface and mantle. These water mass fractions are exceedingly dry (CV: 2.5 wt%, CO: 0.63 wt%) compared to even the driest chondritic meteorites (Wasson & Kallemeyn 1988; Mottl et al. 2007). Similar depletion is seen for carbon in the case of the Earth (McDonough & Sun 1995). All of this is despite water ice being stable at 1 au while the disk was present (Oka et al. 2011). Morbidelli et al. (2016) argues that the inner Solar System is particularly dry due to the presence of proto-Jupiter preventing the drift of volatile-rich material inward from beyond the snow line. Looking to the TRAPPIST-1 system, we see that TRAPPIST-1b and c both likely formed interior to the primordial snow line during the disk lifetime, yet are inferred to contain significant water fractions (Unterborn et al. 2018a). This suggests that volatile-rich material can accrete onto interior planets from beyond the snow line in M-dwarf systems, yet this process seems to not have occurred for LHS 3844b.

The presence of a larger, undetected planet orbiting LHS 3844 that similarly restricted volatile-rich material to LHS 3844b may then explain our results. For example, consider a planet located at the snow line for the \( T = 212 \) K youngest age scenario shown in Figure 14 (2.6 au). Planets with masses in the range of Neptune to Jupiter located at that snow line would induce radial velocity signals with semi-amplitudes in the range 8–143 m s\(^{-1}\) with a period of \( \sim 125 \) days. The radial velocities obtained by Vanderspek et al. (2019) were insufficient to place constraints on the presence of such additional planetary companions in the system, and would require higher-cadence monitoring over an extended period with precision capable of detecting \( \sim 10 \) m s\(^{-1}\) signals. Whether a potentially undetected planet need be Jupiter-mass to create a similar pressure gradient to block inflow from beyond the ice line is thus presently unknown.

Our model excludes magnetic fields, which are typically assumed to protect a planet from atmospheric erosion and therefore effectively lower the erosion rate. The interaction between the stellar wind and a planet’s secondary atmosphere is complex, but has been explored for Earth-size planets with both Earth- and Venus-like compositions (e.g., Dong et al. 2017, 2018, 2020), potentially preserving Venus-like atmospheres on super-Earths for \( >10 \) Gyr. Recent studies, however, have shown that magnetic fields may not help retain an atmosphere at all (Gunell et al. 2018). Whether the presence of a magnetic field would lower the atmospheric erosion rate for LHS 3844b is uncertain. We note, however, that if magnetic fields do substantially lower atmospheric stripping rates, including this effect for LHS 3844b in our models would lower our estimated maximum C budget that is able to produce a \( \sim 10 \) bar atmosphere, by decreasing the atmospheric loss rate, \( E \), in Equation (4). That is, a protective magnetic field would act to favor an even more C depleted planet than we estimate here in order to satisfy the observational constraints.

Our models also consider only primordial heat and radioactive heat production for LHS 3844b. However, additional heat sources could be important. Electromagnetic induction heat is potentially significant for planets on close-in orbits (Kislyakova et al. 2017, 2018), and tidal heating could be significant if LHS 3844b has an eccentric orbit (e.g., Driscoll & Barnes 2015). As Kreidberg et al. (2019b) indicates that the surface of LHS 3844b is solid, any additional heat sources beyond radioactive heat-producing elements must be small enough that they do not result in a fully molten, magma ocean state. The addition of tidal and magnetic induction heat sources at a moderate level, which still result in a mostly solid mantle, would only act to reinforce our main findings. Additional heat sources would hasten the already rapid outgassing of volatiles from the interior to the atmosphere, thus further highlighting that LHS 3844b will likely have outgassed its interior volatile supply to the surface. Moreover, additional heat sources would make the already unlikely case of LHS 3844b having too high a mantle viscosity and too little internal heat to experience significant outgassing even less likely. As a result, whatever volatiles the planet acquired are likely to have been outgassed to the atmosphere. In order for LHS 3844b to have a thin atmosphere today, the volatile abundance must have been low, the rate of atmospheric loss must have been at least an order of magnitude higher than our estimated upper limit based on
Proxima Cen b, or atmospheric stripping by a giant impact must have taken place.

LHS 3844b is considered an ultra-short period (USP) planet, and we placed this system in context with other USP planets by comparing it to 55 Cancri (55 Cnc) and 55 Cnc e. We focus on the differences that could cause LHS 3844b to have no atmosphere, while studies have shown that 55 Cnc e has an atmosphere (Bourrier et al. 2018). The mass for 55 Cnc determined with interferometric and photometric observations by Ligi et al. (2016) is 0.85 ± 0.24 M\(_\odot\), and a rotation period of 38.8 ± 0.5 days was determined by Bourrier et al. (2018). Together, these give a Rossby number of \( R_o = 2.2 ± 0.9\); the large uncertainty is due to the large uncertainty of the stellar mass. Making use of the mass derived by Ligi et al. (2016), which is based upon isochrones (0.874 ± 0.013 M\(_\odot\)), provides stronger constraints on \( R_o = 2.2 ± 0.06\). The Rossby number of 55 Cnc is extremely likely to be significantly higher than that of LHS 3844, meaning that 55 Cnc is a less active star. Additionally, Bourrier et al. (2018) provide a semimajor axis for 55 Cnc e of 0.015 44 ± 0.00005 au, but Vandenspek et al. (2019) give a semimajor axis of 0.006 22 ± 0.00017 au for LHS 3844b. The differences in stellar properties and semimajor axes between these two planets may result in a measurable atmosphere for 55 Cnc e despite a low volatile budget, although detailed observational analysis indicates that 55 Cnc e likely has a magma surface (Demory et al. 2016).

7. Conclusions

The degassing of a planet during the first \(~100\) s of Myr and the subsequent erosion of the atmosphere due to stellar activity are complicated processes with numerous combinations resulting in a variety of outcomes. High stellar activity with low degassing rates can lead to a prevalence of thin secondary atmospheres, while low stellar activity with high degassing rates can result in a retained combination of primary and secondary atmospheres. Investigating the relative contribution of these key processes to atmospheric sustainability and evolution is thus critical for understanding the origin of observed atmospheric masses and abundances. LHS 3844b is a fascinating case in which the atmospheric evolution and influence of stellar activity may be explored in detail. Our age estimate for the host star of \(~7.8\) Gyr provides sufficient time for significant atmospheric loss to have occurred for the planet.

However, whether atmospheric loss over the course of \(~6–9\) Gyr can remove enough mass to explain the currently observed thin atmosphere of LHS 3844b depends on the geological characteristics of the planet. Specifically, our geophysical models of the planet’s thermal and volcanic history indicate that, with a mantle volatile budget similar to Earth’s, \(~200\) bar of CO\(_2\) would be outgassed. Such a large atmosphere cannot be removed by even the fastest stripping rate we consider. We find that a giant impact of a 0.26–0.35 M\(_\odot\) object traveling at the orbital velocity of LHS 3844b could entirely eject this atmosphere, but it would also strip considerable amounts of mantle, resulting in LHS 3844b being an Fe-rich “super-Mercury.” In the absence of a giant impact, LHS 3844b was therefore most likely volatile poor compared to Earth, due to the low partitioning of C into the core of LHS 3844b under both oxidizing and reducing formation conditions. The minimum mantle C concentration we find that is able to produce a <10 bar atmosphere today is \(~10\) times lower than estimates for the Earth. Volatile inventories of this size produce atmospheres of \(\approx 20–25\) bar, which can be stripped to <10 bar over the course of \(~6–9\) Gyr. It is also possible to suppress volcanism and outgassing entirely if LHS 3844b has a high overall mantle viscosity (\(~100\) times more viscous than Earth’s) and low abundance of radioactive heat-producing elements (\(~50\%\) that of Earth’s). However, this unique combination of parameters that can suppress outgassing was rarely seen in our models and also implies a volatile-poor composition for LHS 3844b. These results are consistent with the recent findings of Kite & Barnett (2020), who argued that planets forming close to M-dwarf stars with high planet equilibrium temperatures must have a higher volatile budget than the Earth in order to retain an atmosphere today.

These results imply that LHS 3844b formed both water- and C-poor compared to the Earth. We propose then that it formed interior the the snow line, which provides an upper limit on the degree to which it could have migrated. Furthermore, should planets in M-dwarf systems form via pebble accretion, as in our Solar System, these results suggest the presence of a more massive companion at a larger orbital distance. This companion may have restricted the inflow of volatile-rich material to the inner disk where LHS 3844b formed, in a manner similar to Jupiter’s role in restricting volatile-rich material to the Earth. Early migration and other dynamical changes to the orbit may have resulted in significant modifications to the radiation environment of the planet and subsequent atmospheric loss rates, though these are expected to have occurred early enough to not have played a major role in the overall atmospheric evolution.

The results presented here have been applied to a single case that represents one of the very significant limits on the atmospheric sustainability of a super-Earth planet in close proximity to an M dwarf. With the expectation of numerous further terrestrial planet discoveries around bright stars by TESS (Sullivan et al. 2015; Barchay et al. 2018), there will be additional opportunities to study the relationship between stellar properties and atmospheric evolution. Our results point to the role geoscience can play in contextualizing astrophysical observations, providing key constraints for astrophysical models, and predicting observational tests. As we move toward deeper characterization of individual exoplanetary systems with little directly observed data, these interdisciplinary interactions become even more vital.

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